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TRIASSIC FAULT-LINE DEFLECTIONS AND ASSOCIATED WARPING

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ABSTRACT

Associated with major faults in the Triassic areas in the eastern United States are some well-known warped structures in the relatively downthrown blocks. Much of the warping appears to be related to deflections of the fault lines. Adjacent to salients—fault-line deflections whose outlines are convex toward the dropped blocks—occur anticlinal warps; adjacent to re-entrants occur synclinal warps. The theory is proposed that, during faulting, irregularities of fault surface have caused differential movement, resulting in warping. The salients are *lag points*, where movement of the downthrown blocks has been least; the re-entrants are *advance points*, where movement has been greatest. Tentative explanations are offered for a number of the individual structures. An attempt is made to test the theory by examining part of a Triassic area in which a fault-line boundary has not previously been described. In addition, some data from other areas, outside the Triassic, are included as partially supporting evidence.

INTRODUCTION

Several years ago, through the courtesy of Mr. Kimball Atwood, Jr., of New York City, the writer made an airplane reconnaissance flight over portions of the New England Triassic boundaries. As the eastern boundary was followed northward from New Haven, the Saltonstall and Totoket crescents were seen swinging gracefully in large arcs, with concavity of the crescents facing the Great Fault on the east. In opposition to this curvature the line of the Great Fault itself, over which the plane was flown, appeared in arcs of similar magnitude, facing the trap-ridge crescents in complementary fashion (see Fig. 2, *A*). A causal relationship was suggested which led to the present inquiry, in which the writer has reviewed the

literature dealing with the structure of the (eastern) Triassic areas and has done considerable field work in Massachusetts, Connecticut, New York, and New Jersey. The paper has benefited as the result of many conversations with the writer's associates and especially with M. King Hubbert.

Much has been written concerning the eastern Triassic; but, as far as the writer is aware, the warps have not as yet been satisfactorily explained. One gathers the impression that the majority of geologists who are familiar with these structures attribute them to tensional forces; but there is little evidence in print, as will be shown later, of any unanimity of opinion on the subject. Moreover, the evident relationship between fault-line deflections and warps does not appear to have been previously described. Some of the attempted explanations which follow may seem, on first thought, somewhat forced. It is the writer's belief, however, that the deformational effects that must of necessity be an indirect consequence of the irregularity of some fault surfaces constitute a phase of structural research that has been almost entirely neglected. It is with this point of view that the present paper is offered.

In this paper the terminology dealing with fault movements is employed, except where specific statement is made to the contrary, in a relative sense. Since most of the described warps are in the downthrown blocks, the upthrown blocks are treated, for convenience, as if they had remained stationary during the faulting.

GENERALIZED DESCRIPTION OF WARPS

1. The warps appear on the downthrow side of major faults and die out away from them (see Fig. 1).

2. The warps include anticlines and synclines, pitching toward the fault. The direction of pitch is thus generally the same as that of the regional tilt of the downthrown block, although in some cases a reversal of pitch is present. Most of the warps are "open" structures.

3. The synclines, frequently observed, have been called "half-spoons," "half-boats," "half-saucers," and "plunging folds." Their long dimensions generally parallel the fault, toward which they dip and by which they are apparently truncated. The probable reason

that the synclines have been described more frequently than the anticlines is that they are the larger structures.

4. Within the synclinal warps, strata which are resistant to erosion form surface ridges which show the direction of strike. These are crescentic or horseshoe-shaped in plan view.

5. Along the fault line and within any large crescent there may occur independently some similar but smaller crescents, which terminate against the fault in identical fashion.

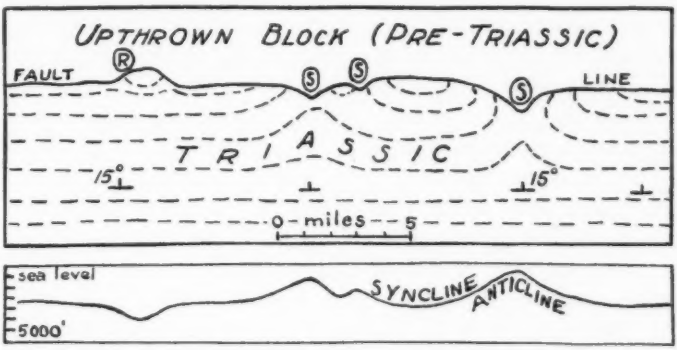


FIG. 1.—Ideal map of a Triassic boundary fault line, showing salients (*S*), a re-entrant (*R*), and associated warping (indicated by structure contours) in the downthrown Triassic sediments. Below is a datum bed in longitudinal profile along the face of the fault.

6. Many of the crescents appear to be directly associated with salients or re-entrants along the fault lines. The general relations are such that a re-entrant along a fault line, concave toward the downthrown block, is confronted in the strata of the downthrown block by a crescent which is concave toward the fault; such a crescent marks a syncline. On the other hand, a salient along the fault line is confronted by an anticlinal crescent, convex toward the fault. Obviously, the direction of concavity of the crescents depends upon the direction of pitch of the warp, which, as noted above, may vary, in which case the direction of concavity of the crescents may vary accordingly. The longitudinal profile of a datum bed along the fault face resembles a series of catenary curves, separated by "nodal points" at the anticlines.

HISTORY OF PREVIOUS INVESTIGATIONS OF THE WARPING

In his classic *Geology of Connecticut*,[1] Percival was the first to call attention to the "curvilinear" form of certain trap and sandstone ridges in the Connecticut Valley lowland and in the Triassic of New York and New Jersey. During the next forty years H. D. Rogers, B. Silliman, Jr., Whelpley, Dana, Wurtz, I. Russell, Cook, Walling, and others inquired into the origin of the crescentic ridges. In general, these early workers assumed an intrusive origin for all of the trap ridges and were thereby led to theories of deformation attributable to intrusion. Walling, however, wrote concerning Mount Toby, Massachusetts: "The strata have been pushed up so as to give them under and around the mountain the form of a basin or canoe. The eastward dips, however, predominate, as might be expected from the general inclination of the basin."[2]

Through recognition of the extrusive nature of some of the trap sheets, Davis was able to demonstrate that many of the crescentic trap ridges and the conformable sandstones above and below them have been faulted and warped. In 1882[3] he wrote the first of a long series of papers emphasizing dynamic deformation of the Triassic of Connecticut. Later[4] he concluded that the warps which are outlined by the crescentic ridges are perhaps in part due to lateral compression in the Triassic rocks; in larger part, however, deformation was transmitted vertically from the basement rocks underlying the Triassic basin. This was accomplished through an ingeniously conceived transfer of lateral compression into vertical movement on a larger number of minute slippage planes. Still later[5] he inferred that the saucer-like folds were formed earlier than, and were truncated by, the boundary fault; differential movement along the fault

[1] J. G. Percival, *Report on the Geology of the State of Connecticut* (published under the direction of the Commissioners appointed by the legislature, New Haven, 1842).

[2] H. F. Walling, "Some Indications of Recent Sensitiveness to Unequal Pressures in the Earth's Crust," *Proceedings Amer. Assoc. Adv. Sci.*, Vol. XXVII (1878), p. 193.

[3] W. M. Davis, "Brief Notice of Observations on the Triassic Trap Rocks of Massachusetts, Connecticut and New Jersey," *Amer. Jour. Sci.*, Vol. XXIV (3d ser., 1882), pp. 345–49.

[4] "The Structure of the Triassic Formation of the Connecticut Valley," *U.S. Geol. Surv. 7th Ann. Rept.* (1885–86), pp. 488–89.

[5] W. M. Davis and L. S. Griswold, "The Eastern Boundary of the Connecticut Triassic," *Bull. Geol. Soc. Amer.*, Vol. V (1894), pp. 525–26.

line could not account for the warping, because the implied difference in throw would be too great. Finally, in 1896,[6] he ascribed the "half-boat-like" basins to irregular warping, with which term he associated no definite quality of deforming process and offered no special explanation. The fault plane was held to be vertical, since he found in the trace of the fault no apparent relation to the curvature of the warped structures.

In New Jersey, according to Kümmel,[7] the large warp displayed in the Watchung Mountain crescent had been formed prior to faulting.

During the next twenty-five years little was written concerning the origin of crescent ridges or their half-boat extensions underground. In 1922, as the result of a detailed examination of the Triassic fault boundary in southern Connecticut, W. L. Russell[8] differed with Davis on several significant points. First, since he found that the warps "all have the same relations to the Great Fault, and all die out rapidly west of it, they appear to be connected with its formation." Second, he admitted the possibility of unequal subsidence along the fault plane as an explanation for the warps. Third, he believed the fault plane to dip 30°–60° toward the downthrown block. Fourth, he pictured the irregular boundary fault as a single, perhaps modified, line or zone of displacement, rather than a series of unrelated fractures. He appears to have accepted Davis' opinion that the trace of the fault bears no significant relation to the pattern of warping.

In a brief paper Foye[9] mentioned the anticlinal fold east of Cedar Mountain in southern Connecticut and "others known within the valley," suggesting that they are the result of compressional strains propagated across the valley from the west, "dammed" at the boundary fault upon encountering the rigid crystalline block.

[6] "The Triassic Formation of Connecticut," *U.S. Geol. Surv. 18th Ann. Rept.*, Part II (1896–97), p. 86.

[7] H. B. Kümmel, "The Newark System," *New Jersey Geol. Surv. Ann. Rept. State Geologist*, Part II (1897), p. 78.

[8] "The Structural and Stratigraphic Relations of the Great Triassic Fault of Southern Connecticut," *Amer. Jour. Sci.*, Vol. IV (5th ser., 1922), pp. 489–97.

[9] W. G. Foye, "Abnormal Dips near the Eastern Boundary Fault of the Connecticut Triassic," *Science*, n.s., Vol. LIX (1924), p. 240.

Recently Bain, commenting on observed structures at the southern horn of the Greenfield–Deerfield–Bull Hill crescent, wrote: "Structure of the Triassic of the basin is therefore in no wise due to faulting in or along the edge of the basin after extrusion of the Deerfield sheet."[10] From this it follows that Bain does not consider post-depositional faulting to be directly related to the half-spoon of the Deerfield trap sheet.

EXAMPLES OF DEFLECTIONS AND WARPS

SOUTHERN CONNECTICUT

An ideal type locality at which to start an investigation of the Triassic warps is southern Connecticut, not only because the continuous presence of three interbedded lava flows permits reliable structural mapping but because the fault patterns which cut them are, on the whole, simple. That is, there are two main fault systems: the marginal Great Fault and the diagonal intralowland faults (see Fig. 2, *A*). Although it is undoubtedly true that either system is more complex than its representation on the map, and that there may be interfingering at places, still it is possible to differentiate the two systems.

The map of the southern Connecticut Triassic in Figure 2, *A*, suggests that the warps were formed earlier than the diagonal faults. Therefore, if it be assumed as a working hypothesis that the warps are directly dependent for their formation upon the Great Fault, then a reconstructed map, with the diagonal fractures "unfaulted," should reveal any earlier relationship which may have existed between the Great Fault and the warps. In making a reconstructed map of this type (Fig. 2, *B*) the writer follows at *D* the map given by Russell,[11] who shows the line of the Great Fault extending a small distance farther west than it is located on Davis' map.

With the diagonal faults thus omitted, it is evident that there are three large crescents formed by the trap ridges of the lowland. These may be called the Saltonstall, Totoket, and Higby crescents, named after the prominent ridges formed by the main trap sheet.

[10] G. W. Bain, "The Northern Area of Connecticut Valley Triassic," *Amer. Jour. Sci.*, Vol. XXIII (5th ser., 1932), p. 71.

[11] *Op. cit.*, p. 485.

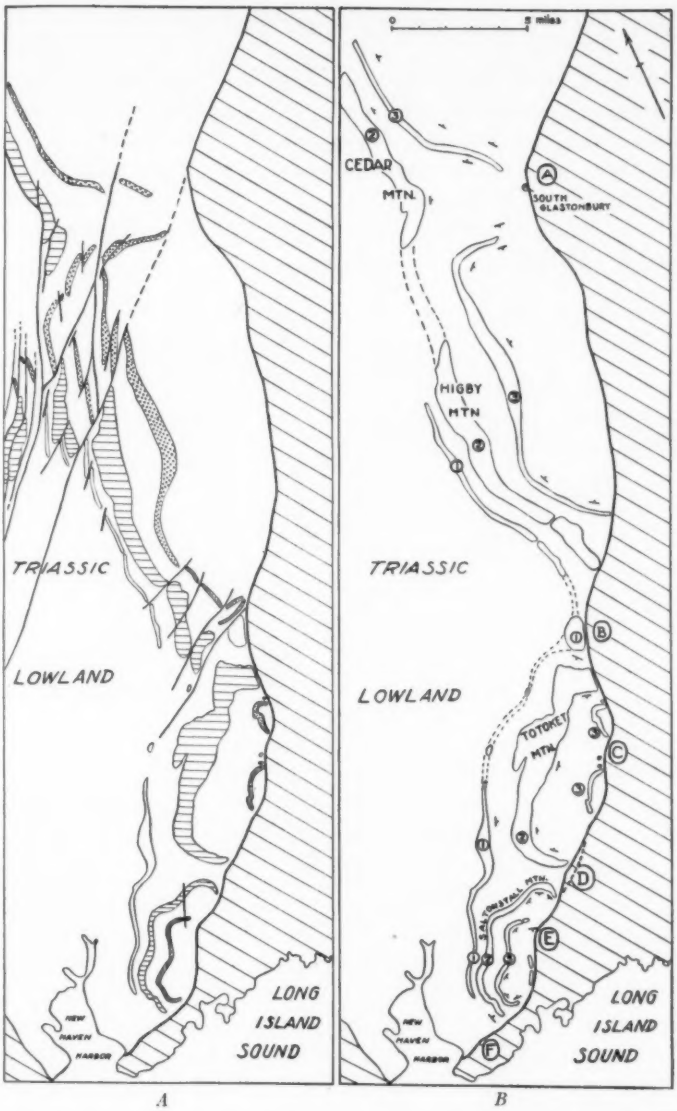


FIG. 2.—*A*, map of a portion of southern Connecticut (after Davis, Longwell, and Russell) showing Crystalline Upland (oblique lines) and Triassic Lowland. Within the lowland are three trap sheets, shown with different patterns. *B*, reconstructed map of the same area, with intra-lowland faults omitted.

The Great Fault appears as an irregular line comprising a number of large and small arcs whose concavity faces west. Roughly, there seem to be three larger arcs, from *A* to *B*, from *B* to *E*, and from *E* to *F*. The second of these is interrupted by two salients at *C* and *D*.

As pointed out by Russell, the warps in the sediments and interbedded lavas die out west of the Great Fault. For example, oppo-

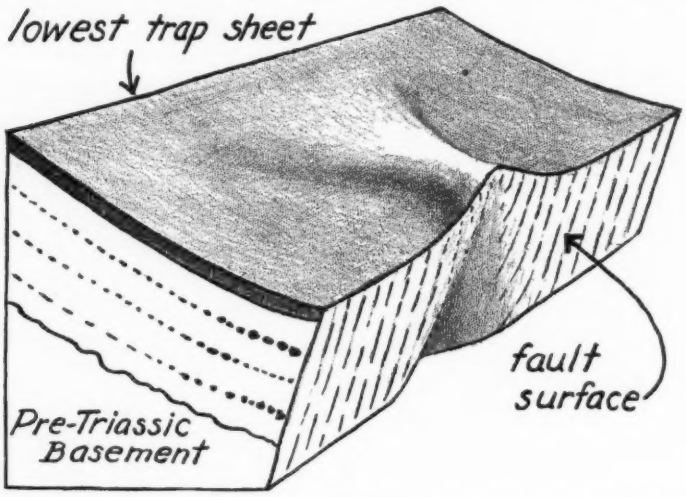
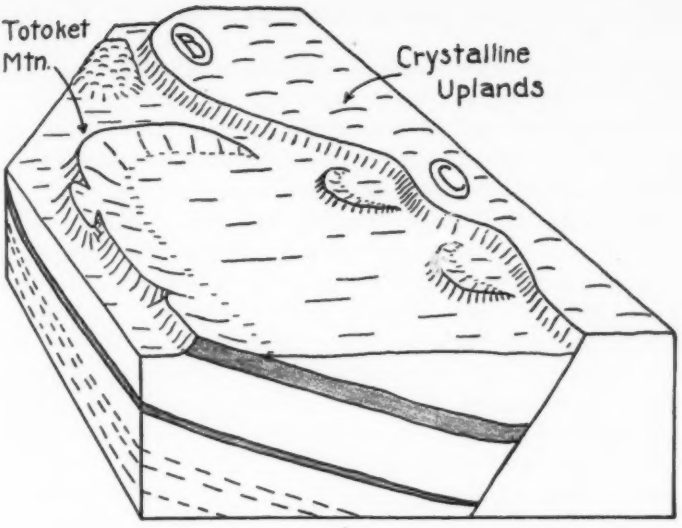


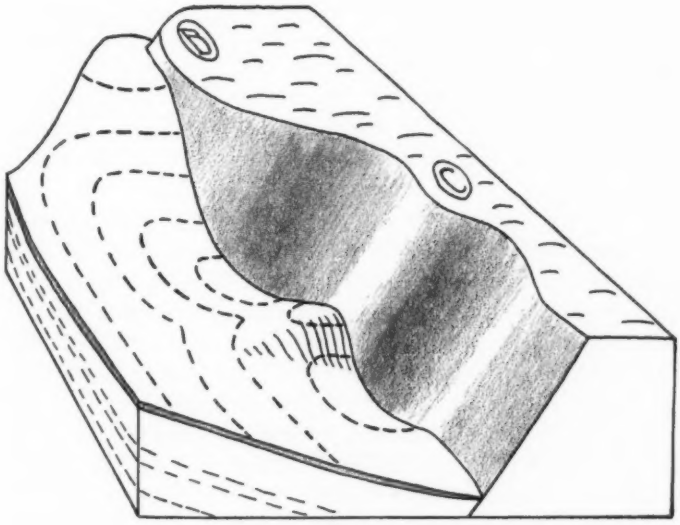
FIG. 3.—Idealized diagram showing deformation of the lowest trap sheet opposite salient *D* of Figure 2, *B*. The salient is pictured as growing larger with depth.

site *D*, the outcrop of the western (lowest) trap sheet is only slightly curved in the same warp that causes the main (middle) sheet to outcrop in recurved horns. The surface of the lowest sheet at this locality is shown in idealized form in Figure 3.

A more complex relationship is shown in the block diagrams of the Totoket crescent (Figs. 4, *A* and *B*). A small patch of the lowest sheet appears just west of the Great Fault opposite *B*. As shown in Figure 4, *B*, this sheet decreases in elevation along the face of the fault about 4,000 feet within a horizontal distance of 3 miles. It was this rapid change of elevation which led Davis to reject the theory of differential throw. In the light of present knowledge of Basin Range



A



B

FIG. 4.—*A*, block diagram of part of the Totoket Mountain area. *B*, the same, with all Triassic sediments above the lowest trap sheet removed.

faulting, this objection cannot be considered valid. For example, Gilluly,[12] in describing a Basin Range fault with a displacement of 3,000 feet, points out that in one direction from this measured displacement the fault dies out completely within 2 miles. Powell, Dutton, Gilbert, and others have described similar cases.

The configuration of the lowest trap sheet (Fig. 4, *B*) reveals several other significant relations. Below the salient at *C* is an upward bulge of the dropped block; the result is that contours on the surface of the trap sheet show crescentic patterns. Hence, if a horizontal section be taken through the dropped block at the elevation of any given contour, the truncated trap sheet will then have an outcrop pattern identical with that of the contour. The actual crescentic outcrops of the trap sheets in the valley lowland form patterns very similar to the hypothetical contours.

It is a short step from the inference concerning the shape of the concealed portion of this trap sheet to the hypothesis that any salient along the line of the Great Fault marks a lag point at which the downthrown block has been held back in its descent along the fault surface. This generalization is supported by the structural relations at *A*, *B*, *C*, *D*, and perhaps *E* and *F* of Figure 2, *B*. As a corollary, it might be suspected that at the larger salients would be found the more effective lag points. This suspicion is only partially substantiated by the comparative magnitude of the salients and warps.

NEW JERSEY AND PENNSYLVANIA

The descriptions which follow of warping in the lowland of New Jersey and Pennsylvania are necessarily oversimple. The structural ensemble in this area is more complex than is the case in Connecticut. With the exception of the Watchung trap sheets in northern New Jersey, good marker beds do not occur adjacent to the boundary faults. In the southern part of the area, crosscutting of sediments by intrusive dolerite sheets makes it difficult to trace out with precision the warping of the Triassic strata.

The simplified map of part of the New Jersey–Pennsylvania Triassic (Fig. 5) shows that the boundary (largely a fault line) between

[12] J. Gilluly, "The Geology and Ore Deposits of the Stockton and Fairfield Quadrangles, Utah," *U.S. Geol. Surv. Prof. Paper 173* (1932), p. 83.

the downthrown Triassic and the highlands is an irregular line. If the broken central portion of the line can be thought of as a salient zone, many of the characteristics previously enumerated in the generalized description of deflections and warps are found here. Two complementary crescents are outlined. That on the west, of which only one horn is shown, may be called the "Hunterdon crescent," since the Hunterdon Plateau in New Jersey closely follows its strike. On its western end it is interrupted by a fault. The other

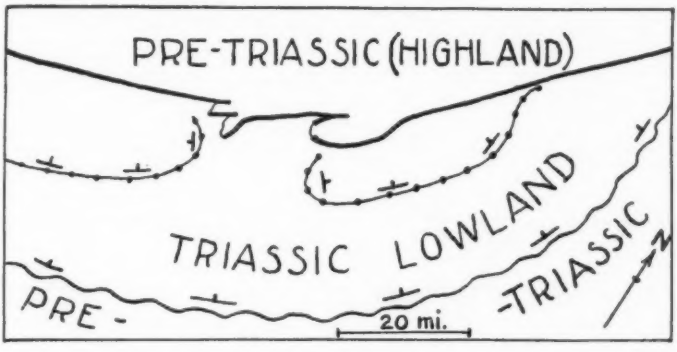


FIG. 5.—Generalized map of part of the New Jersey–Pennsylvania Triassic, showing fault-zone boundary (heavy line), depositional boundary (crenulate line), and two large crescentic warps (dotted lines).

crescent, on the east, is the Watchung crescent, named after the trap ridges which form it. Twenty miles south of the fault zone boundary there is no trace of these large warps. They are further discussed in a later paragraph.

On the inner side of the Watchung crescent pronounced warping occurs along the Ramapo fault line (Fig. 6). The outstanding features at this locality are: the Ramapo fault, with several gently curving deflections; the long crescent formed by the first and second Watchung trap sheets; and, as indicated by the sinuous outline of the third trap sheet, sharp warping between the crescent and the fault. For the termination, or horns, of the large Watchung crescent the present theory suggests the possible existence of footwall protuberances at depth; should these not be actually present, differ-

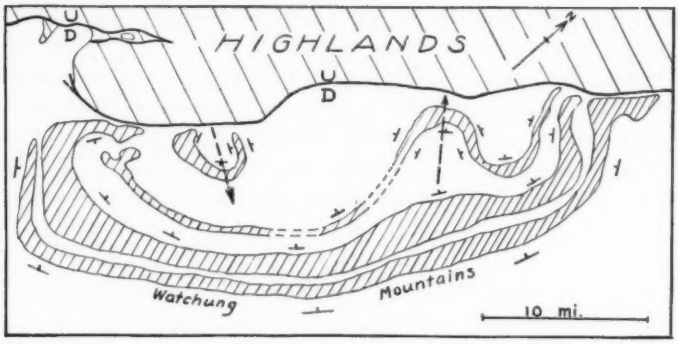


FIG. 6.—Map showing warping in the New Jersey Triassic Lowland adjacent to the Ramapo Fault. (After *New Jersey Geol. Surv. Atlas Sheet 40*, Kümmel revision [1931].)

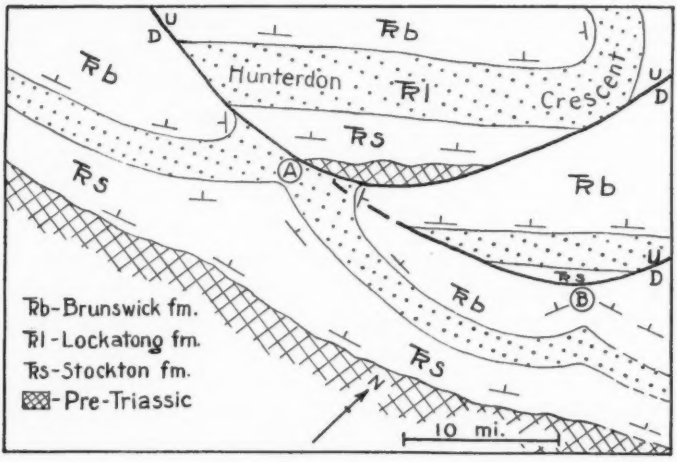


FIG. 7.—Map of part of the New Jersey–Pennsylvania Triassic, showing relation of two faults to warping. Intrusive dolerite omitted. (Slightly generalized from Pennsylvania and New Jersey State Geologic Maps, 1931 editions.)

ential subsidence and drag, the cause being unknown, provides an explanation that is serviceable, although incomplete, for the large-scale warping. The smaller warps, in close proximity to the fault, are apparently connected with configuration of fault surface in the same way that the warps within the Totoket crescent, Connecticut, are connected with the Great Fault in that area. Anticlines occur opposite the salients; synclines are opposite re-entrants. The two anticlines pitch in opposite directions, as shown on the map by arrows indicating the pitch of the anticlinal axes. This is taken to indicate that, where the plunge is outward from the fault, the drag has been greater. Although the deflections are gently curving in outline, and although the warps do not "fit" exactly with the corresponding deflections, the theory already proposed to explain similar warps in Connecticut may be applicable here.

A somewhat different type of warping appears in connection with the faults shown in Figure 7. Because the large Hunterdon crescent is cut on two sides by the long curved fault (the Flemington-Furlong-Chalfont fault), it appears possible that here, as in the Connecticut area, there were two periods of faulting: the early boundary deformation, in which the Hunterdon crescent was produced, and the later intrabasin faulting, shown in Figure 7. There is no direct evidence, however, that the warping at *A* and *B* was produced in the second period; and it is possible that all the deformation took place at about the same time, with slight overlapping.

On the hypothesis that the warp at *A* is at a lag point, causally related to the adjacent fault, how can the location of the lag point be explained? One explanation is pictured in Figure 8, where the curved fault dies out at both ends. Since the maximum friction would occur at the point of greatest throw, the warp (i.e., drag) should also be found there. At first sight, this explanation might appear satisfactory; it is simple and, in addition, is supported by the known dying-out of the large fault at its western end.[13] Opposed to this is the fact that at its eastern end, where it joins the highland boundary near Cushatunk Mountain, New Jersey, the amount of

[13] F. Bascom, E. T. Wherry, G. W. Stose, and A. I. Jonas, "The Geology and Mineral Resources of the Quakertown-Doylestown District, Pennsylvania and New Jersey," *U.S. Geol. Surv. Bull. 828* (1931), p. 43.

throw must be roughly 11,000 feet,[14] equal to that at the middle curvature of the fault (*A* of the diagram), 25 miles to the southwest. Another objection is that whereas, according to this explanation, the dip of the strata in the upthrown block should exceed that of the beds in the downthrown block, actually this is not the case.

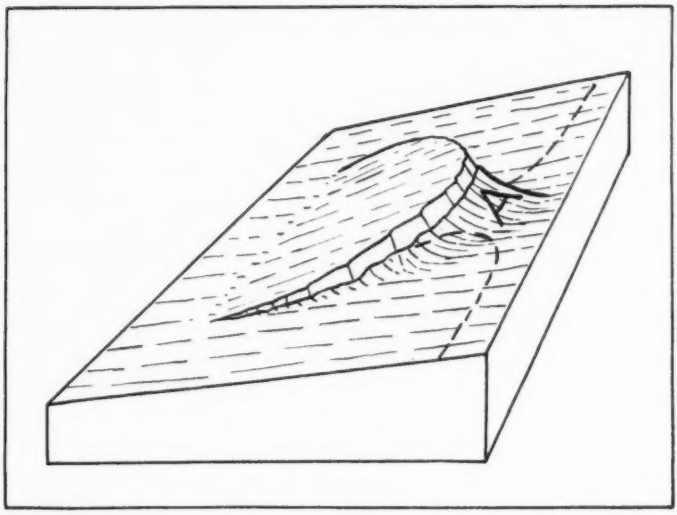


FIG. 8.—Ideal restoration of crescentic fault of Figure 7, showing deformation in the relatively downthrown block as resulting from differential drag.

An alternative explanation is shown in Figure 9, which is a schematic analysis of the curving fault surface. On the assumption that the fault surface dips uniformly and that the fault is normal in type (evidence that most of the large Triassic faults are normal will be presented later), a plan view of this surface, projected upward from the outcrop line of the fault (*PMNS*), would reveal three surface elements: *OMPR*, *ONST*, and *OMN*. Of these, the first two are approximately plane surfaces; the third is the surface of a quarter-cone. Any beds descending along the two plane surfaces would show no differential drag, but beds which descend the face of the conical wedge must undergo distention. The distention would in-

[14] Kümmel, *op. cit.*, pp. 85–86.

crease with depth, and the natural result would be an increase in drag.

As a consequence of the theory that the deformation at *A* is dependent on the fault, the question arises: Why is the assumed differential drag found only in the hanging wall? Why is there not

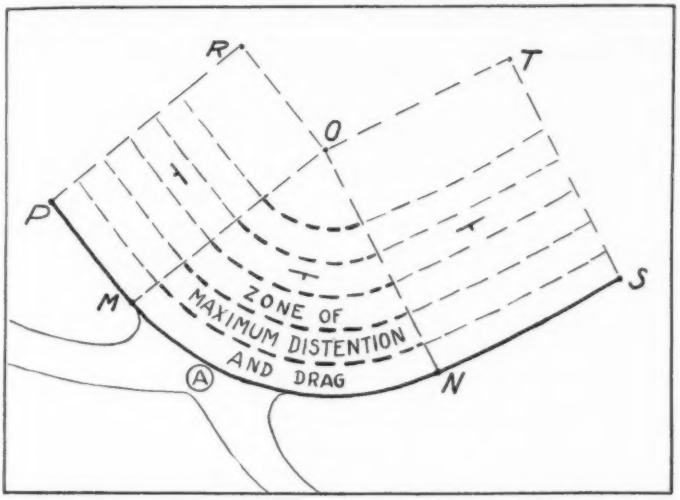


FIG. 9.—Contoured projection upward above the present erosion surface of an ideal extension of the surface of fault shown in Figure 7. Actual outcrop of fault: *PMNS*. It is assumed that the fault is normal and that the fault surface dips in the direction shown by dip symbols.

a similar deformation in the adjacent footwall? The answer may lie in the fact that pressure exerted downward by the hanging wall upon the footwall meets the resistance of the entire crustal thickness below, whereas pressure upward against the hanging wall is resisted by a comparatively thin crustal layer and deformation is more easily accomplished. Or it may be that differential competence of the rocks involved accounts for their different behavior.

Admittedly, this explanation involves the unproved assumption that the fault is a normal fault. Also, the Flemington-Furlong-Chalfont fault actually comprises several faults and is therefore more complex than shown. Its throw is not the same from end to end. Hence, the location of the warp at *A* and similarly of the lesser warp

at *B* may be fortuitous and unrelated to the curvature of the adjacent faults. Probably the chief value of ideal reconstructions such as Figure 9 is in indicating the effects which one might hope to find duplicated on fault surfaces of the type postulated.

THE SAND BROOK AND NEW GERMANTOWN CRESCENTS

In the foregoing examples of warping, emphasis has been placed upon bulges occurring as anticlines at the salients. Another type of warp, which may also be dependent upon the configuration of the face of the fault, is exemplified by two small crescents near Sand Brook and New Germantown, New Jersey (Fig. 10). The re-entrants with which they are associated have been selected as type examples of advance points—points at which the hanging wall has descended notably farther than elsewhere along the fault. The crescents, aptly described by Kümmel as "horseshoes," terminate abruptly against faults.

> These two small areas . . . . have preserved remnants of trap sheets that were doubtless once much more extensive. Each is but the eastern spoon-shaped end of a syncline that has been cut off by a fault to the west, and each contains remnants of two separate masses of trap of exactly similar forms and relations. The larger mass in each case outcrops in a westward-pointing crescent and curves downward beneath the shales within. The smaller caps a rounded hill between the points of the crescent, thus resting upon the sediments that overlie the synclinal sheet below.[15]

Lewis demonstrated that, although the crescents are 15 miles apart, they are probably contemporaneous and parts of the same sheets. How does it happen that in both instances the warps abut against re-entrants along the fault lines? Kümmel writes on the structure of the Sand Brook crescent:

> On the southeastern side of the fault, also, there are some interesting examples of truncated structures. The best example of this kind is the syncline . . . . whose axis is inclined northwestward. The trend of the fault is nearly transverse to the arms of the syncline and the folded structure is very prettily terminated by the fracture. The small extrusive trap-sheet near Sand Brook is included in this fold and the northern limb of the curve seems to be cut off by the fault.[16]

[15] J. V. Lewis, "The Origin and Relations of the Newark Rocks," *New Jersey Geol. Surv. Ann. Rept. State Geologist*, Part III (1906), pp. 115–16.

[16] *Op. cit.*, pp. 83–84.

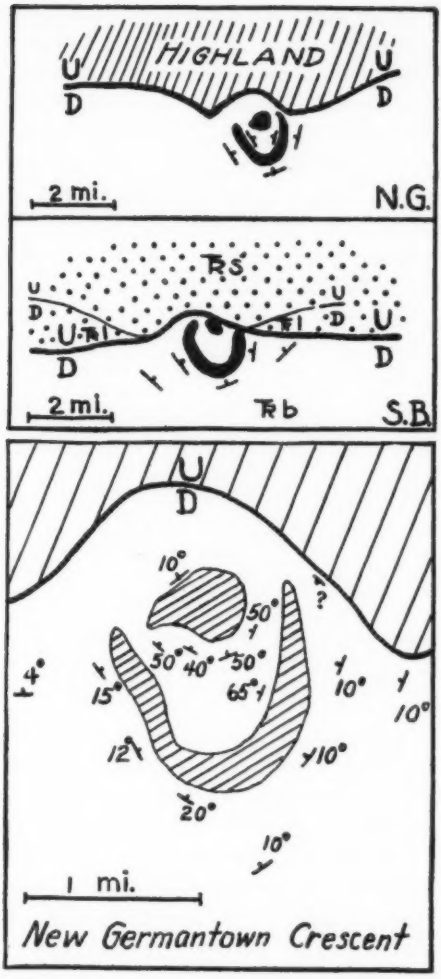


FIG. 10.—Maps of the New Germantown (*N.G.*) and Sand Brook (*S.B.*) crescents (after Kümmel). Below: The New Germantown crescent on a larger scale (after Darton). On atlas maps New Germantown now is known by the name "Oldwick."

The theory advanced in the present paper is that many—perhaps most—of the warped structures in the Triassic basins, where located adjacent to faults, are not "truncated" by the faults but are instead the *result* of the faulting. The block diagram in Figure 11 suggests that at the point where the dip of a normal-fault surface is locally steepened the strata of the descending block encounter less frictional resistance and hence may descend to a lower level along the face of the fault than the beds which move downward along the adjacent, more gently dipping surface. This block might serve as an ideal restoration of the Sand Brook structure.

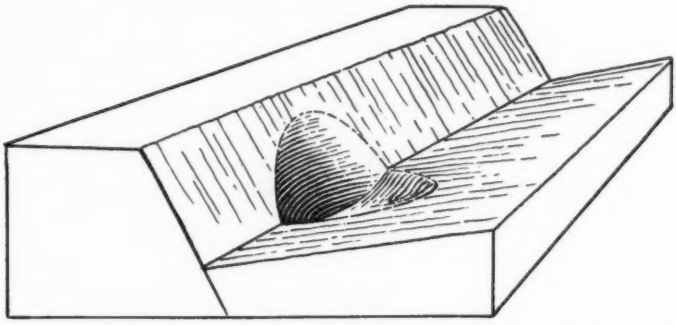


FIG. 11.—Hypothetical fault deflection, showing a steeply dipping re-entrant, the site of an advance point in the downthrown block.

Such a conception automatically brings to mind several questions. What lies underneath? Is it implied that the downthrown area of "advance" is underlain by subcrustal magma? Or that the relatively down-dropped Triassic blocks finally came to rest "floating" on subcrustal material, receiving partial support from the footwalls of the fractures? In the writer's viewpoint such questions go much farther than the theory just presented. The explanation proposed for this (and for New Germantown) structure postulates an actual "gravity" fault. If the presence of such faults in nature be permitted, it is not necessary to know the exact nature of the crustal support. It seems almost self-evident that, once a gravity fault is inaugurated, the footwall must contribute some support. From this it follows that, if at any point this support is diminished, the drop-

ping mass would, of necessity, subside farther than adjacent areas, regardless of what may lie beneath.

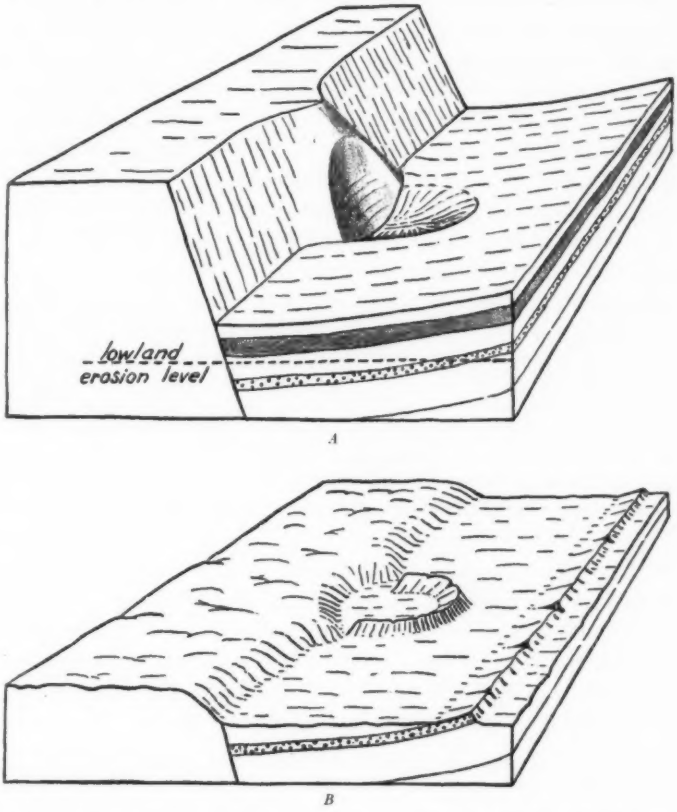


FIG. 12.—*A*, a steeply dipping concavely surfaced fault slice is pictured as responsible for the location of the advance point in the downthrown block. *B*, the same block is shown after erosion.

A modified form of this explanation is proposed for the New Germantown warp (Figs. 12, *A* and *B*). The outline of the fault at this locality permits the hypothesis that prior to the development of the re-entrant there was a salient of the crystalline highlands pro-

truding into the lowland block. It may be supposed that the added pressure against the salient portion of the footwall as the result of distention in the downsliding block produced an auxiliary steeper shear along the face of the salient. The sheared-off footwall spur thus detached has presumably descended sufficiently far to be concealed at depth. Evidence suggesting that increased pressure upon salients may be followed by slicing along their faces is given by Gilbert in his description of the western fault boundary of the Wasatch Range. He enumerates five salients which are portions of the fault line convex toward the intermont basin; of these, four are associated with sliced-off spurs; at the fifth, a spur is described but is interpreted as a landslide. In his summary Gilbert writes:

> These relations suggest that the local conditions that determined the spurs were the same as or connected with the conditions that determined the salients. Irregularity in the fracture surface of the great orogenic fault was occasioned by heterogeneity of the faulted terrane. Elements of irregularity in the direction of strike are manifested in flexures of the fault line, and to this class belong the salients. Co-ordinate with these flexures are irregularities on lines of dip, and as the fault movements were in the direction of dip there was a constantly renewed discordance between the forms of hanging wall and footwall. Also, because the weight of the overlapping valley block enforced contact, there was a constantly renewed adjustment.[17]

Gilbert's observations on deflections were confined almost entirely to the upthrown blocks. In the eastern Triassic basins, because of denudation through several cycles of erosion, it is seldom possible to observe details of the upthrown blocks; on the other hand, many of the structures of the downthrown block are directly ascertainable, and therein lies their value in a study of this sort.

DATA ON FAULT-LINE DEFLECTION AND WARPING IN SOUTHERN MASSACHUSETTS

From Rockville, Connecticut, at the eastern Great Fault boundary, the escarpment of the crystalline uplands runs northward in a straight line to Chicopee River in Massachusetts. This 20-mile stretch presumably marks a fault line along which the total throw may be as great as it is farther south in Connecticut—16,000 feet, according to Russell's minimum estimate. Viewed from the valley

[17] G. K. Gilbert, "Studies of Basin-Range Structure," *U.S. Geol. Surv. Prof. Paper 153* (1928), p. 32.

lowland, this wall readily conveys the impression that it is a fault-line scarp. Yet there is little mention in the literature of any actual evidence for faulting. Davis says of the Connecticut portion:

> . . . . The northern part of the border has few neighboring Triassic outcrops. The evidence of faulting there is less distinct than further south; yet on making a general section on true scale across the Triassic lowland, it is difficult to account for the disappearance of the strata unless cut off by a marginal fault.[18]

Two miles north of Scantic (see Fig. 13) a quartz vein, 50 feet wide, appears to have come up along the contact between the Triassic sandstone and the pre-Triassic rocks of the upland. For a quarter-mile the vein dips almost uniformly 48° westward. On the side bordered by the sandstone it contains many red patches which may be replaced fragments of Triassic sediments. Thin sections were made of the patches; but if they are, indeed, remnants of sediments, replacement by quartz has been too thorough to permit their definite identification. As far as the evidence may be interpreted, it suggests that the boundary at this point is a normal fault.

At Wilbraham, $2\frac{1}{2}$ miles south of Chicopee River, Emerson[19] has given evidence that is entirely acceptable for a fault boundary. North of Wilbraham, as far as the Holyoke Range, the boundary has not been described as a fault line. That it is faulted is suggested by the abrupt eastward swing of the Holyoke trap range into the re-entrant of the boundary line at Three Ponds. In order to check this surmise and to check the theory now under consideration, the writer made a reconnaissance along the boundary from Chicopee River around the protruding Bagg Hill salient and the Three Ponds re-entrant on the north. According to the theory developed in this paper, if the contact were a faulted one, there should occur in the Three Ponds re-entrant crescentic warping associated with the fault in a predictable pattern.

At Lyons Pond there is an abrupt termination of a series of northeastward-striking hills of Triassic sandstone and conglomerate. The line of hills ends on the northeast in a clifflike slope overlooking a narrow swampy valley which separates the Triassic from the crystal-

[18] Davis and Griswold, *op. cit.*, p. 523.

[19] B. K. Emerson, "The Geology of Old Hampshire County," *U.S. Geol. Surv. Mono. 29* (1898), pp. 360–61.

line uplands. The dip of the conglomerate beds immediately adjacent to this border valley is 40° southeast. The writer considers

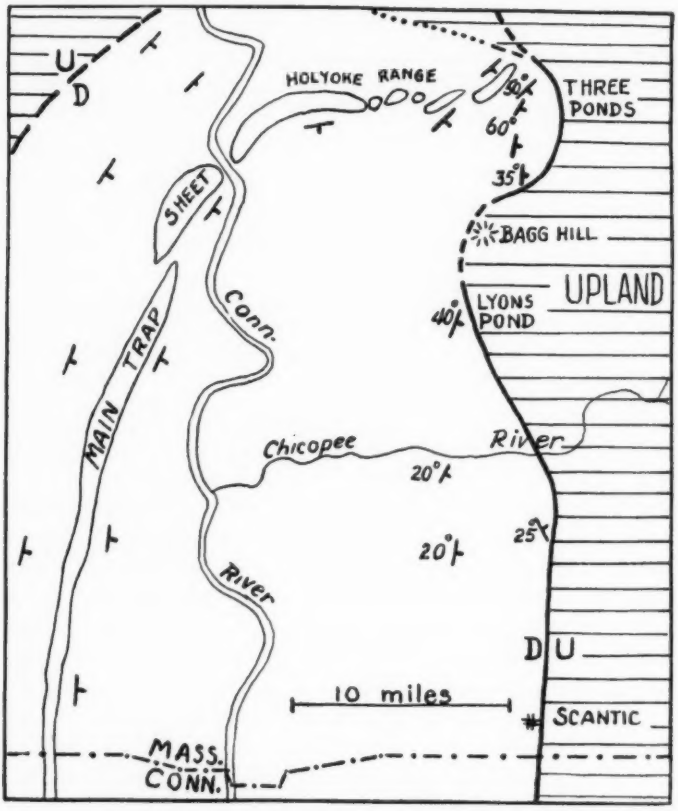


FIG. 13.—Map of a portion of the Triassic in southern Massachusetts, showing relation of warping to the Great Fault on the eastern border. Lined area: pre-Triassic upland. Note the varying strike of the Triassic at the Three Ponds re-entrant.

this to be compelling evidence that the boundary in this section is a fault line and that the valley is a fault-line valley.

Around the salient and the recurving boundary to Three Ponds, softly contoured crystalline hills meet the lowland with a low, subdued scarp. Footing the scarp, there generally occurs a depression,

which is commonly occupied by a stream or a swamp. As shown on the map (Fig. 13) the strike of the sandstones forms a distinct curve whose concavity faces eastward. The physiographic features, the strong dips toward the crystalline uplands, and the actual occurrence of the anticipated crescentic strike pattern of the Triassic, shifting 30° within the Three Ponds re-entrant, are taken to constitute evidence indicating a fault boundary.

SUGGESTIVE DATA FROM OTHER AREAS

An attempt to compare Triassic structures with deformations associated in other areas with deflections of normal faults has been only moderately successful. Several requirements should be met. In the first place there must be present a normal fault of considerable length, containing deflections along an uninterrupted line of fracture. In the second place the fault must be located in an area that is otherwise undeformed. That is, it must be reasonably certain that there has been only a single deformational episode; otherwise the possibility of "cross folds," "later orogenic movements," and the like will vitiate any conclusions. In order to satisfy this requirement, the ideal situation would be that of a large area containing flat or gently dipping strata—except, of course, adjacent to the faulting. Difficulties in finding such an area (outside of the Triassic) have been, first, to find faults that are, beyond any question, normal faults and, second, to find places where the dragged strata next to such faults are contoured structurally.

These requirements are at least partially satisfied in the Mexia and Tehuacana fault zones of Texas, which have been so admirably described by Lahee.[20] Only a few structural features, as summarized by Lahee, will be given here. There are a number of normal faults dipping on the average between 40° and 50°, with total displacements ranging from 250 feet to more than 600 feet. Adjacent to the faults, on the upthrow side, oil has accumulated in gently deformed anticlinal areas 1,500–36,000 feet in length, which are clearly associated genetically with the faults. The maximum structural relief of any pool is 175 feet; the minimum, 12 feet. Toward the ends of

[20] Frederic H. Lahee, "Oil and Gas Fields of the Mexia and Tehuacana Fault Zones, Texas," *Structure of Typical American Oil Fields*, Vol. I (Tulsa: Amer. Assoc. Petr. Geol., 1929), pp. 304–88.

several of the pools there is a tendency for the fault surface to assume a steeper dip. The better-developed folds are broadest and highest near places where the fault intersection with the oil-containing bed makes a bend and generally where the fault has its least average angle of inclination. The displacement of the major fault seems to be at a maximum in the middle part of each field, with a decrease in throw toward the extremities.

These features are believed by Lahee to accompany a regional torsion, and the main folds are ascribed by him to the compressional forces involved in the torsion. The present writer finds the evidence for regional torsion compelling, but that the folds themselves are the effect of a compressional component of torsion is not equally clear. In any case, it may be noted that these structures are unquestionably associated genetically with the faults and that there appears to be a relationship between deflection and folds. Because only the upthrow sides of the faults have been oil-bearing, data are lacking on structure in the downthrown blocks. Oddly enough, if the entire crustal area of some of the Mexia faults were inverted, the faults would, of course, still appear normal; but the anticlines would appear as synclines, appropriately located in complementary fashion to the salients, which in the inverted form would be re-entrants.[21] In other words, the actual folds in the relatively upthrown blocks appear roughly as "mirror images" of the Triassic structures of the downthrown blocks, as described in this paper. To the writer this points forcefully to the inference that deformations along normal-fault deflections occur in predictable pattern in both upthrown and downthrown blocks.

On deductive grounds it would appear possible that irregularities on the surfaces of thrust faults ought also to produce recognizable deformation in the blocks involved. Should such deformation prove to be different in type from that accompanying normal faults, then the relationships between deflections and attendant deformation might serve, in the absence of other evidence, as a criterion for discriminating between normal faults and thrust faults. The practical difficulty in many cases would be to distinguish between structures produced by the irregularity of the fault surface and any unrelated

[21] See, e.g., Figs. 18 and 27 of the paper by Lahee.

structures, such as often occur, for example, on both sides of a thrust fault in an area that has undergone considerable compression.

In addition to the suggestive relationships in the Mexia zone, there appears, in a description of transverse faults at Kennecott, Alaska, some rather striking evidence that irregularities along fault surfaces do actually produce noticeable effects. In discussing fracturing at Kennecott, Lasky proposes the hypothesis that

> transverse disturbances found stopping against a main fault are dependent on the presence of flexures along the fault surface; that they occur in the moving block and indicate that block by their presence; that the transverse fracture is often indicative of the kind of flexure and movement that caused it; and that a certain definite relationship of strike and dip exists between the transverse fracture and the fault flexure.[22]

Although this is not an exact parallel to the Triassic warping, the theory as stated by Lasky may well be considered to have anticipated the fundamental principle of the present paper. Lasky's inference that the presence of transverse structures indicates the moving block would be a very important one if it could be demonstrated that this is generally the case. Substituting "warping" for "disturbances" and applying the theory to the Triassic basins, it would mean that most of the major fault movements have been the result of actual sinking movements of the relatively dropped blocks. The writer, while not questioning the possibility that Triassic deformation may have been of a "sinking" nature, does not think that any evidence yet in hand warrants the statement as fact. Moreover, as has been pointed out, deformations in the Mexia fault zone are genetically associated with fault-line deflections, yet appear in the upthrown block; and it seems unlikely in this area that the relatively upthrown blocks have been the active ones.

OTHER HYPOTHESES TO ACCOUNT FOR WARPING

DEFORMATION FROM INTRUSION

One of the first theories to explain the Triassic warping was that the sediments have been deformed by the intrusion of dolerite dikes and sills. Beyond doubt the intrusions have been directly respon-

[22] S. G. Lasky, "Transverse Faults at Kennecott and Their Relation to the Main Fault Systems," *Amer. Inst. Min. Eng. Tech. Pub. 152* (1928), p. 17.

sible for some deformation, although it is difficult to estimate its magnitude. But this hypothesis was almost entirely ruled out when Davis showed that many of the trap sheets are extrusive rocks and hence may not be utilized to explain deformation of the sediments with which they are interbedded. Of the warped areas observed by the writer, the only cases in which this explanation might be employed are in Pennsylvania, where intrusive dolerite sills and dikes occur in the zone of warping. As the intrusive sheets appear to be warped together with the sediments, it seems probable that the intrusions are not responsible for much of the observed warping.

DIFFERENTIAL INITIAL DIPS AND COMPACTION OF SEDIMENTS

Dynamic deformation is excluded in the hypotheses that would explain major Triassic structures by differential initial dips and compaction of sediments. In early investigations of the Triassic, initial dips were frequently invoked to account for the attitude of the strata. Included within the folds are extrusive basalt sheets whose original attitude must have been nearly horizontal. Proof of faulting and deformation of the sediments having been furnished by Davis, the "initial-dip" theory was left with little supporting evidence.

In accounting, however, for structures along the eastern border of the Triassic in Massachusetts, Bain[23] places great emphasis on the effects of consolidation of sediments. The present writer[24] has briefly indicated the grounds for his belief that this eastern Triassic boundary is determined by lines of postdepositional faulting. Since Professor Bain has been kind enough to go over in the field with the writer some of the critical exposures, these will be discussed in more detail in a separate paper. In the writer's judgment consolidation effects have been relatively unimportant; it is believed that the Deerfield trap sheet has been warped and faulted, together with underlying and overlying beds, subsequently to deposition. Whatever may be the case in Massachusetts, it is clear that initial dips and dips resulting from consolidation will not serve as a general explanation for the warping in Triassic areas because of the high dips involved in demonstrable relationship to faults at many places.

[23] *Op. cit.*, p. 77.

[24] Girard Wheeler, "The West Wall of the New England Triassic Lowland," *Conn. Geol. and Nat. Hist. Surv. Bull. 58* (1937), p. 71 and Pl. I.

DEFORMATION FROM LATERAL COMPRESSION

Foye suggested that compressional stresses transmitted across the Connecticut lowland were "dammed" upon reaching the massive highland block. This theory does not satisfactorily explain the warps, because the direction of pitch of the structures is parallel with the direction in which the assumed compression would have been operative.

It is not known whether the warps are the result of compression or tension. But since the anomalous structures die out with increasing distance from the faults, we must accept the conclusion of W. L. Russell that the warping is in some way connected with the formation of the faults. This is certainly the case in southern Connecticut and in all probability in many of the warps in the other basins. The writer's view[25] is that the stress affecting the Triassic beds was almost entirely tensional, although some torsional effects—which, of course, include compression—may well have been produced. In the present paper simple tension has been emphasized, although it is recognized that to isolate a structure as the product of tension alone is to run the risk of oversimplification.

WARPING PRIOR TO FAULTING

Although it is fairly clear that the warps are somehow connected with faulting, it is not so evident that they were formed after movement took place on the fault. Some of them have been considered as pre-fault structures, truncated by the fractures which followed. Davis and Kümmel employed this conception as a criterion of fault boundaries. An objection to the truncation theory is that so many of the folds appear cut off in identical fashion—the typical "half-boat" form is duplicated again and again. A greater variety of forms ought to occur if pre-existing structures had been truncated. Also, the truncation theory does not explain why the axes of the warps trend at right angles to the faults and why the warps are so commonly related to deflections.

DIFFERENTIAL SUBSIDENCE WITHOUT REFERENCE TO CAUSE

It has been stated in the preceding discussion that differential subsidence (Fig. 14) may be the only available explanation for some of

[25] *Ibid.*, pp. 69–73.

the warps, as, for example, the large Watchung and Hunterdon crescents. But it should be kept in mind that fault surfaces are seldom plane surfaces. A protuberance on a fault surface may be concealed at depth and hence not reveal itself at the present surface of erosion by a deflection of the fault line. Along a straight fault line above such a protuberance, an anticline in the strata of the dropped block may be the only evidence of its location. On the other hand, where a visible relationship exists between deflections and warps, it is obvious that a more comprehensive explanation for the differential movement is desirable.

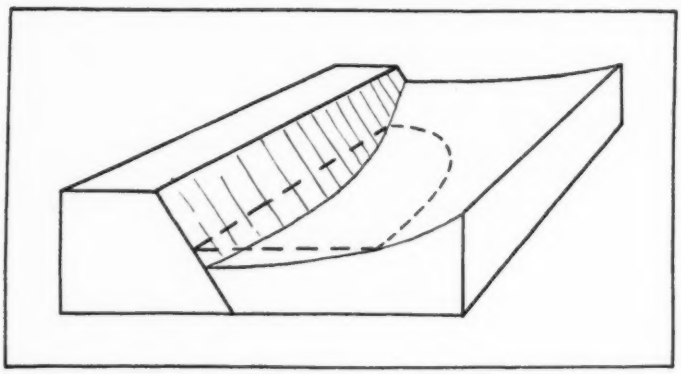


FIG. 14.—Warped structure resulting from differential subsidence, the cause being unknown.

OBJECTIONS TO THE LAG POINT THEORY

THE DIP OF THE FAULTS

A fundamental assumption of this theory is that the Triassic faults described are normal faults. Since the validity of this assumption may be questioned, it is necessary to state the available evidence.

1. Although Davis believed, on deductive grounds, that the plane of the Great Fault in Connecticut is nearly vertical,[26] at the only place where the fault plane was exposed, near Manchester, Connecticut, he found that the dip is 55° westward, making it a normal fault[27] at that locality.

[26] "The Triassic Formation of Connecticut," *op. cit.*, p. 126.

[27] Davis and Griswold, *op. cit.*, p. 526.

2. In a tunnel section through Saltonstall Ridge in southern Connecticut, a short distance from the Great Fault, Longwell observed sixteen minor faults.[28] All are normal; the dip of the fault planes averages 67°. Longwell suggests that these are "sympathetic" faults and that, if this be the case, the Great Fault is of the normal type.

3. From a study of the Great Fault in southern Connecticut, Russell[29] concluded that the fault is normal and that the fault plane dips from 30° to 60°.

4. The dip of the quartz vein at the eastern boundary near Scantic, Massachusetts, previously mentioned in this paper, is 48° toward the lowland. This suggests that the fault plane also dips in the same direction and that the fault is therefore normal.

5. The dip of a similar quartz vein on the eastern side of the lowland near Mount Toby, Massachusetts, is 45°–70° westward toward the lowland. As far as the writer could determine, at least part of this vein, like that near Scantic, is located at the boundary of the Triassic. In the same vicinity the dip of the slickensided hanging wall of the quartz-barite vein at the Leverett Lead Mine is 64°–79° northwest. One-half mile west of this mine, dips toward the boundary as high as 25° are common in the Triassic. The data given above are believed to be strongly suggestive of a normal-fault boundary.

6. On the western boundary of the Massachusetts Triassic at the Loudville Lead Mine the footwall of an opening on a quartz-barite vein shows a vertical fluting which dips 60° toward the lowland. The vein has ascended between walls of Triassic sandstone and pre-Triassic granite. These relations indicate that the boundary is marked by a normal fault.

7. At the Bristol Copper Mine in Connecticut the western border of the Triassic is formed by a normal fault; the fault plane dips 70°. Mineralization has taken place along this fault.[30]

8. Little has been written on the dip of the faults in New Jersey and Pennsylvania. In an early paper Kümmel says: "About 75

[28] C. R. Longwell, "Notes on the Structure of the Triassic Rocks in Southern Connecticut," *Amer. Jour. Sci.*, Vol. IV (5th ser., 1922), pp. 223–36.

[29] *Op. cit.*, p. 490.

[30] A. M. Bateman, "Primary Chalcocite; Bristol Copper Mine, Connecticut," *Econ. Geol.*, Vol. XVIII (1923), pp. 122–66.

faults are known to occur. . . . . With but few exceptions, all the known faults are reverse faults."[31] Later, however, he writes: "The [Triassic] beds are cut by many nearly vertical normal faults, the amount of dislocation varying from a few inches to several thousand feet. . . . . Uplift accompanied by northwestward tilting and normal faulting . . . . brought up again the crystallines."[32] A large majority of the minor faults observed by the present writer in New Jersey and Pennsylvania are normal faults.

9. According to Wherry,[33] most of the Triassic faults observed in Pennsylvania are high-angle normal faults.

Normal Triassic faults have been described at several places in the southern basins; but since warping in the same areas has not been clearly pictured, they will not be cited as evidence in that connection. In summary, although there is some absolute evidence that the boundary faults of the Triassic basins are normal, their designation as such, having been arrived at in part by inference, must still remain tentative.

FAILURE TO ACCOUNT FOR SOME OF THE WARPS

A second objection to the theory is that it fails in some places to account for known warps, and that in others the "fit" is not perfect. The suspicion arises that the location of some of the anticlines and synclines is accidental and that any relationship existing between deflections and warps is only an apparent one. Such a suspicion is well founded, for example, in considering the Watchung and Hunterdon crescents in New Jersey (Fig. 5); for, although, as already pointed out, the general setting for these crescents is similar to that for the smaller salients and warps, it is difficult to picture the "salient zone" in this case playing the same role during deformation that the smaller salients are assumed to have played. The scale in the case of these larger crescents may be too large for the hypothesis based upon small-scale examples.

There are several answers to this objection. First, the theory does not demand that all basin warping is dependent upon the con-

[31] "The Newark Rocks of New Jersey and New York," *Jour. Geol.*, Vol. VII (1899), pp. 43–44.

[32] "Geologic Section of New Jersey," *ibid.*, Vol. XVII (1909), p. 366.

[33] E. L. Wherry, oral communication.

figuration of the fault surfaces. Any fault surface is more complex than pictured in the accompanying illustrations. Inequalities of relief on the fault surface and of the forces acting at depth, as well as variations in lithology of the faulted blocks, are several of the many factors which may be responsible for the observed deformations.

Second, in so far as some of the warps do not fit exactly with the deflections of the fault lines, it should be kept in mind that, when dealing with such large faults, whose throws have been estimated by many observers to be from 6,000 to 16,000 feet, it would be surprising, rather than otherwise, if the trace of every one of the warps were in perfect accord with the hypothetical patterns at the present erosion level. Just as soon as the lower part of a salient is concealed from view beneath the surface of a lowland, our conception of its form is inferential and therefore subject to error. Concerning the complex conditions under which any extensive fault takes place, Gilbert says:

> . . . . The surface of eventual fracture is neither plane nor of simple curvature. As soon as movement takes place the walls cease to be fitted the one to the other. A prominence of one wall leaves the companion hollow and becomes opposed to an incompatible surface. Adjustments are then necessary to preserve contact. . . . . It is easy to understand that details of configuration may depend on elements of structure not revealed at the surface.[34]

THE FORM OF THE SALIENTS

The admission of ignorance in the preceding section is logically followed by a third objection: Why should it be assumed that the form of a salient is such as to produce unusual drag in the hanging wall? Where the form of the salient is conic, convex toward the dropped block, and the base of the cone is at depth, it may not be difficult to visualize distention and drag in the downsliding block. The deformation caused by other salient forms is not so obvious. For example, what deformation would take place if the salient has the form of a half-cylinder, whose axis inclines in the same direction and with the same dip as the fault plane on either side of the salient? (see Fig. 4, *B*). After fracture occurs, so long as gravity plays a part in the movement of the dropping block, there might well be a tendency for part of the material in the hanging wall above the

[34] *Op. cit.*, p. 21.

cylinder to slide downward and sideways off its crest. This is because the direction of true dips on an inclined cylinder are (in every place except the crestline) away from the crestline and have a magnitude greater than the pitch of the cylinder. The sideslipping tendency might produce distention and drag in somewhat the same fashion as occurs with a conic salient.

In preliminary clay experiments made by the writer the half-cylinder salient has actually produced an anticlinal warp at the salient. Further experiments along similar lines are at present being carried out in the Rutgers University geology laboratory by Mr. H. Jobbins. However, a complete analysis of the mechanical principles which would determine the shape of salients and re-entrants along normal-fault lines has not yet been made. The deflection forms shown in the illustrations are therefore open to the criticism that they are speculative. They have been chosen to bring out in purely suggestive form the causal relationship which appears to exist between deflections and folds.

Another objection of similar nature might be that the long irregular fault lines described in this paper are not the continuous lines of single fractures that they appear to be but are instead fracture zones made up of a number of short connecting faults. Considering the curving fault front of the west side of the Wasatch Range, Boutwell writes:

> It seems that the mountain-making fault did not raise the eastern mass over the western mass by a single uniform movement all along its length but let the western mass sink piecemeal in separate parts while the eastern mass rose more continuously; thus the sinking mass headed against the rising mass in concave bights from 1 to 10 miles across. The trace of the fault as followed along the mountain mass today is therefore neither rectilinear nor extremely irregular but is composed of a considerable number of entrant curves meeting in blunt cusps.[35]

If we employ Boutwell's conception in the case of the Triassic faulting, the sharper salients might be pictured as representing "keels" between crescentic faults. Now, if a geometric construction be made of the forms that would underlie the fault trace, it will be found that, provided the dips on the fault surfaces remain ap-

[35] John M. Boutwell, "Wasatch Front," *XVI Int. Geol. Cong. Guidebook 17 (The Salt Lake Region)* (1933), pp. 34–35.

proximately uniform in amount, the inclination of any keel must be less than that of the fault surfaces on both sides of it. The keel might therefore be expected to retard the movement of the crustal mass sliding downward over it. However, if the constant inclination of the fault surfaces be maintained, each keel ultimately disappears with depth and in its place a smooth conic salient appears. At this

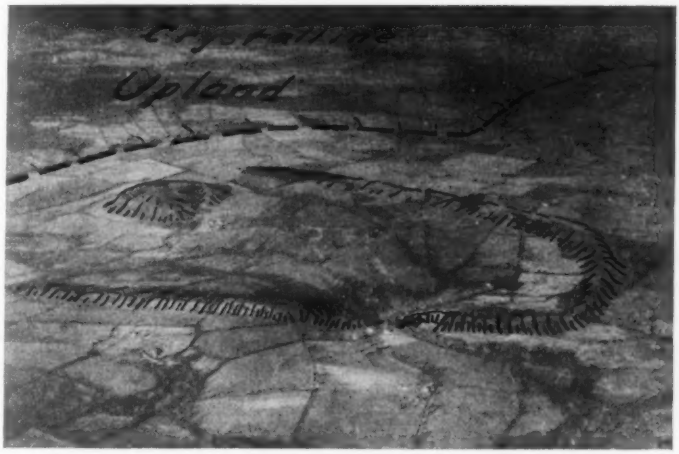


FIG. 15.—Airview (retouched) of New Germantown warp. The view is an oblique one, looking northward, taken at 4,500 feet above the lowland. The crescentic ridge is the outcrop of a spoon-shaped trap sheet which dips inward beneath a second trap remnant, showing as a small hill. Dips are as high as 65°. The area included in the crescent is about 1 square mile.

point it will be observed that, if any crescentic re-entrant be limited on each side by conic salients whose bulk becomes larger with increasing depth, then the re-entrant must become correspondingly smaller; the hanging wall, upon descending, would be wedged into the re-entrant and hence undergo compression at right angles to the fault line. Compression of this sort may have taken place at some points on the Triassic faults. For example, the strong dips (35°–60°) in the Sand Brook and New Germantown (Fig. 15) synclinal warps are suggestive of compression. But even if it should be demonstrated that some compression is present as an accompaniment of normal

faulting, there would still remain to be explained the fact that within these concave re-entrants the downslid strata have descended so much farther than at other points.

The writer does not believe that the constant dip utilized in thus constructing the curvature of fault surfaces is found in nature. Indeed, there are observational data for many normal faults that stand in contraversion to such a conception.[36] Hence, unless a geometric analysis of fault surfaces has a demonstrable relation to the actual mechanics of deformation, either from a theoretical point of view or on the basis of observed data, it can have, as stated above, only a suggestive value. But a more factual obstacle to the "multiple crescentic fault" conception, as applied to the Triassic faults, appears to be that some, at least, of the scallop patterns represent zones or lines of fracture, adjacent to which large crustal masses in the hanging wall have moved essentially as units (see parts of the border fault line in Fig. 2, *A*). On this account the writer is inclined to follow the opinion of W. L. Russell, as expressed for part of the Connecticut border fault, and to expand this conception to include much of the irregular eastern Triassic border faulting. The faulting is pictured as representing lines or zones of displacement rather than series of short separate fractures. It is true that modifications and offsetting faults may be seen at places, but this is only a normal expectation where a heterogeneous crust is deformed during intermittent movements.

CONCLUSION

Much of the warping in the Triassic areas of the eastern United States has a definite relationship to fault-line deflections. It appears possible that the warping may be the result of differential drag on uneven fault surfaces. Whatever the actual cause, it is possible, once the line of one of these faults is known, to predict the type of warping that occurs in the downdropped blocks.

Theoretically, it should be possible to predict the probable deformation effects dependent upon the irregularities of any known fault surface.

[36] See, e.g., J. Gilluly, "Basin Range Faulting in Utah," *Bull. Geol. Soc. Amer.*, Vol. XXXIX (1928), p. 1106.

THE HOWELL STRUCTURE, LINCOLN COUNTY, TENNESSEE[1]

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ABSTRACT

A small area of highly disturbed, contorted, and brecciated strata centering at Howell, in north-central Lincoln County, Tennessee, has been mapped and studied in detail. The salient structural feature is a circular area of intensely deformed Black River and Trenton rocks, which have been uplifted approximately 100 feet above their normal positions. This circular area is composed of jumbled blocks of limestone imbedded in a matrix of shatter breccia. The major deformation is believed to have been post-Trenton and pre-Fernvale in age. Overlying the shattered strata is the Fernvale formation, the relative thickness and lithology of which point directly toward deposition in a graded crater. The Fernvale shale unit shows moderately high dips. This partial disturbance of the Fernvale possibly indicates a mild renewal of the forces which caused the major deformation. The younger Silurian and Mississippian formations are relatively undisturbed. A magnetic survey indicates a "closure" of 335 gammas, $1\frac{3}{4}$ miles southwest of the center of the disturbance. While no conclusive evidence has been observed to support either the cryptovolcanic or the meteoritic hypothesis of origin, the Howell structure is considered tentatively as an example of the cryptovolcanic structures as interpreted by Bucher.

INTRODUCTION

During the latter part of 1937 the writers mapped about 10 square miles in the vicinity of Howell, Lincoln County, Tennessee, studying in detail a circular, intensely deformed area about 1 mile in diameter that is centered at Howell. This area is in the north-central part of Lincoln County and is about 6 miles north-northwest of Fayetteville (Fig. 1). It is readily accessible, as the Lewisburg-Fayetteville highway (State Highway No. 50) crosses the center of the area.

So far as is known, the late J. W. Young, of Fayetteville, was the first to take notice of this small area of intricate structure. He showed it to several visiting geologists and in 1934 gave its location to the present writers. No detailed map, however, was made until 1937, when the writers completed their map (Fig. 2).

[1] Published with the permission of the State Geologist, Tennessee Division of Geology.

PHYSIOGRAPHY

North-central Lincoln County is in the highly dissected Highland Rim region of south-central Tennessee, where large streams, such as

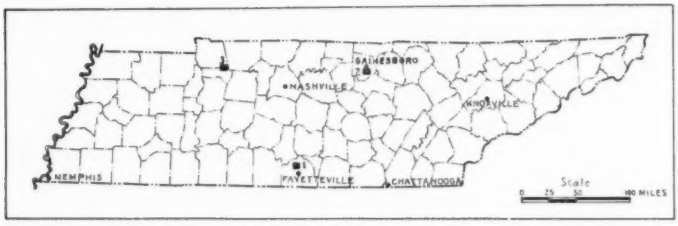

FIG. 1.—Index map of Tennessee, showing the location of (1) Howell structure, Lincoln County; (2) Flynn Creek disturbance, Jackson County; and (3) Wells Creek basin, Stewart and Houston counties.

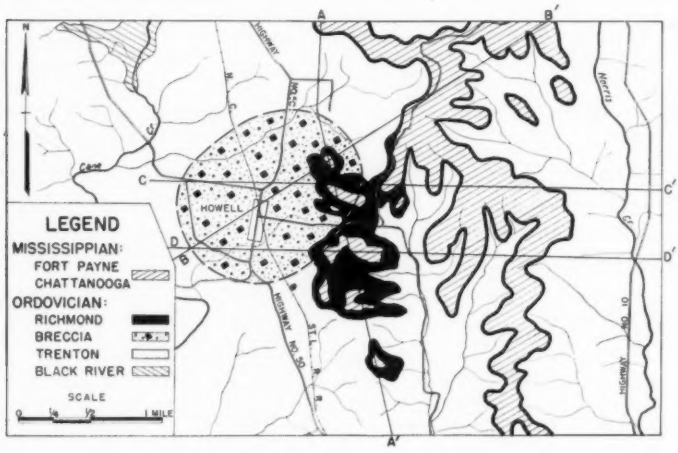

FIG. 2.—Areal geologic map of the Howell structure, Lincoln County. The heavy broken line indicates the approximate border of the circular area of intense brecciation.

Cane and Norris creeks, have cut wide valleys practically to the level of the Central Basin. Separating these wide valleys are narrow interstream ridges that represent remnants of the former more extensive Highland Rim level. Much of the valley level averages about 700 feet above sea-level (Howell is 748 feet above sea-level), while the ridges have an average elevation of about 1,035 feet.

STRATIGRAPHY

The consolidated rocks exposed in the area range from Black River age (Ordovician) to Osage age (Mississippian), inclusive. The formations are briefly described, as follows:

ORDOVICIAN SYSTEM

MOHAWKIAN SERIES

BLACK RIVER GROUP

Carters limestone.—The Carters limestone, which is the oldest formation exposed in the area mapped, occurs in the northwestern corner of the map along Cane Creek. Here the rocks dip southward below creek level within a short distance of the edge of the map (Fig. 2). In the small creek that flows westward across the highway (State Highway No. 50) just north of Howell, large blocks of Carters limestone occur, but here they have been uplifted appreciably above their normal position.

Both the upper and lower members of the Carters limestone occur in this area. The upper member is a thinly bedded, dove-colored, fine-grained limestone, about 12 feet in thickness, and contains specimens of *Tetradium cellulosum* and ribbon-like bryozoans. The topmost part of the lower member, which is all that is exposed of this member, is a massively bedded, gray, fine to medium or coarse limestone that contains chert lenses and specimens of *Columnaria halli*, *Tetradium columnare*, *Streptelasma profundum*, and *Stromatocerium rugosum*.

TRENTON GROUP

Hermitage formation.—This formation, which consists of blue argillaceous limestone and calcareous shale which weather brown, is exposed along parts of Cane Creek. It also occurs in the small creek that flows westward across the highway (State Highway No. 50) just north of Howell. About 1,000 feet east of the highway bridge over this creek, blocks of the Hermitage formation vary from steeply dipping and vertical to overturned. Here it has been uplifted, for it normally should not be exposed at the elevation of that locality. The Hermitage formation contains the usual specimens of *Dalmanella fertilis*. No thickness was measured, but Bassler[2] reports 52

[2] R. S. Bassler, "Stratigraphy of the Central Basin of Tennessee," *Tenn. Div. Geol. Bull. 38* (1932), p. 77.

feet of Hermitage near Belfast, which is only 15 miles north of Howell.

Bigby limestone.—The Bigby is a granular, cross-bedded, phosphatic limestone exposed at several localities within the area mapped. No thickness was measured. Bassler[3] reports this formation to be 33 feet in thickness at Columbia, Maury County, some 40 miles northwest.

Cannon and Catheys formations.—These two formations constitute more or less of a unit and are considered together. They consist of a series of blue, fine to medium-grained limestones that have a total thickness of approximately 200 feet. They are widely exposed in the area mapped, occupying the higher parts of the stream valleys and most of the slopes of the high ridges that divide drainage between Cane Creek and Morris Creek.

BRECCIA

Breccias composed of angular fragments of limestone occupy a circular area 1 mile in diameter and centering around Howell. The fragments range in size from shot up to large blocks many feet in dimension and occur in a matrix of powdered limestone. Much of this breccia consists of small, angular to subangular fragments the size of walnuts. Within this type of shatter breccia occur large angular blocks of limestone that may or may not be brecciated. Many of these blocks of limestone are cut by dikelike stringers, or veins, of injected powder breccia, which suggest forceful intrusion along fractures while the injected material had a "mushlike" consistency. Good exposures of the breccia, including large blocks of Hermitage and Carters formations, occur upstream along the small creek that flows westward across the highway (State Highway No. 50) north of Howell.

The breccia contains blocks and fragments of Carters, Hermitage, Bigby (undoubtedly, although none was actually identified), Cannon, and Catheys formations. It is overlain nonconformably by Fernvale shale and limestone and is therefore of post-Catheys, pre-Fernvale age.

[3] *Ibid.*, p. 83.

CINCINNATIAN SERIES

MAYSVILLE GROUP

Leipers formation.—No exposures of this formation were found in the area mapped; but, as it is believed that the Leipers formation originally covered much, or all, of central Tennessee,[4] the limestone probably was present here but subsequently removed from this part of Lincoln County by pre-Richmond erosion.

RICHMOND GROUP

Arnheim formation.—No exposure of this formation was found, and it is not known whether it is absent owing to erosion or to nondeposition in this part of Lincoln County.

Fernvale formation.—The Fernvale apparently occupies its normal horizon throughout the area mapped, for, although it may be absent locally, such localities were not found. In this area the Fernvale formation consists of four units, described, as follows, from oldest to youngest.

1. Green, calcareous siltstone, containing *Dalmanella meeki:* This nonresistant unit is often not exposed, owing either to local absence or to being covered by the blanket of chert rubble from the higher Fort Payne chert. It was measured at several localities, and its known thickness varies between 6 and 12 feet.

2. Coarsely crystalline limestone: This unit is very widespread, having been found at localities where exposures could be expected to occur in spite of a soil covering and the rubble of Fort Payne chert. It consists of irregularly bedded, massive, coarsely crystalline limestone that is often red or green in color. Throughout much of the area it averages 10 feet in thickness, but at the head of the small creek north of Howell it is as thick as 35 feet. Here it rests directly upon the breccia, the siltstone zone with *D. meeki* being absent. At this locality the Fernvale limestone unit is itself in part brecciated. But this brecciated Fernvale contains fragments only of Fernvale limestone imbedded in a matrix of limestone of an identical lithology. It likewise does not contain fragments of the underlying breccia (Trenton and Black River) but maintains a distinct unity as a

[4] C. W. Wilson, Jr., "The Pre-Chattanooga Development of the Nashville Dome," *Jour. Geol.*, Vol. XLIII (1935), p. 479.

veneer of Fernvale limestone covering the underlying greatly disturbed, shattered, and jumbled breccia. The Fernvale limestone unit contains beds of unbrecciated limestone and may be traced laterally into Fernvale limestone showing no brecciation. For these reasons it is believed that the brecciation of the Fernvale limestone unit was in part contemporary brecciation that occurred shortly after the partial consolidation of some of the beds of Fernvale limestone. This probably occurred as the result of readjustment in the underlying jumbled breccia of Trenton and Black River strata.

The presence of *Rhynchotrema capax*, *R. dentata*, *Sowerbyella clarksvillensis*, etc., definitely fixes its age as Richmond.

3. Shale unit: This unit is usually greenish- to grayish-brown shale or mudstone and may contain zones of red shale. The predominant textural type is silt, but it contains local streaks of rounded silt and quartz pebbles up to the size of peas, though averaging about the size of shot. Layers of hard siltstone and fine-grained sandstone also occur at different horizons. These layers have about the same color as the shale and mudstone with which they are interbedded. This unit contains many ferruginous zones that weather brown and red. A few crinoid stems and brachiopods found in the shale unit indicate its marine origin.

The thickness of the shale unit varies greatly. East of Howell it was measured up to 95 feet. It maintains a thickness greater than 40 feet throughout the area of widespread Richmond outcrop, shown on the map (Fig. 2) a short distance east and southeast of Howell, and through which the line of section, A–A', is drawn from north to south. Throughout the remainder of the area the shale is poorly exposed, but thicknesses of 10–15 feet were measured. It is possible that the unit is absent locally, permitting the Mississippian strata to rest directly upon the persistent Fernvale limestone unit.

4. Lenticular sandstone units: At most localities the overlying Chattanooga shale rests on the Fernvale shale unit. Locally, however, thick lenses of sandstone occur, cutting down into the shale unit in a way suggestive of stream channeling. This sandstone varies from medium to coarse grained and may contain rounded pebbles of quartz as large as shot. It is usually dark brown in color, owing to its ferruginous content.

About 2,000 feet south of the stream divide on the road slightly over half a mile east of Howell, the sandstone unit is well exposed and has a thickness of 25 feet. At the northern end of the small outlier of Mississippian, just northwest of where the east-west road crosses the drainage divide slightly over ½ mile east of Howell, are several exposures of local sandstone lenses. Another exposure of the sandstone unit is reached by driving eastward 1 mile from Howell, turning south with the road for ¼ mile, and climbing to the middle of the southern slope of the large eastward-pointing spur. Here is exposed what is probably a planoconvex lens of sandstone 70 feet wide and about 12 feet in thickness in the middle. The sandstone lens cuts down into the Fernvale shale unit and is covered by Mississippian strata, but with occasional intervening remnants of Brassfield chert.

No fossils were found in these lenticular bodies of sandstone. Their distribution and shape suggest stream channeling following the regression of the shallow sea in which the underlying shale unit had been deposited.

SILURIAN SYSTEM

MEDINAN SERIES

ALBION GROUP

Brassfield limestone.—No exposures of this formation were found, but at two localities concentrations of residual chert containing species of typical Brassfield corals were found. These blocks of chert, which are characteristic of the Brassfield limestone, occur along the side of a hill at a horizon beneath the Chattanooga shale and above the Richmond shale and sandstone units. It is believed that they represent erosional remnants of a formerly more extensive Brassfield limestone, remnants so thinned by erosion that the limestone was completely leached during pre-Chattanooga erosion, leaving only the more resistant chert.

One of the two localities is around the small outlier of Mississippian just northwest of where the east-west road crosses the drainage divide slightly over ½ mile east of Howell. The other is reached by driving eastward 1 mile from Howell, turning south with the road for ¼ mile, and climbing along the southern slope of the large east-

ward-pointing spur (the first one south of the east-west road from Howell).

MISSISSIPPIAN SYSTEM

Kinderhook Series

Chattanooga shale.—This formation is only about 5 feet in thickness, and consequently exposures are rare. Where seen, however, it consists of typical black, carbonaceous shale. The Hardin sandstone member, which forms the basal unit of the black shale, is usually a dark brown, phosphatic sandstone.

Osage Series

Fort Payne chert.—The highest ridges and hills in the area mapped are capped with Fort Payne chert, which varies in thickness from a few feet up to 50 feet or more. Unweathered Fort Payne is rare, and practically all of the formation here has had most of the calcareous material leached, leaving layers of blocky, brown chert. These blocks of chert have rolled down the steep hills and cover much of the outcrop belts of older formations.

Structure

REGIONAL SETTING

The Howell area is situated on the southern flank of the westward swing of the Nashville dome. Regional dip, therefore, is to the south and is at an average angle of considerably less than 1°. Recent structural data[5] have indicated the presence in this region of a series of small folds with northwest strikes that are superimposed upon the flank of the Nashville dome. Many of the anticlines are asymmetrical, having a gently dipping flank and a much steeper, and possibly locally faulted, flank. Dips on the gentle flank average 2° or 3°, while those on the steeper flank may be 5° and 10° or locally higher.

The Howell disturbance, characterized by a small circular area of highly disturbed, contorted, and brecciated strata, is located near the crest of an anticline in the series of folds with northwest strike; but this relationship is believed to be fortuitous. Two other areas of localized, intense deformation are known in central Tennessee. One

[5] Wilson, "Stones River and Black River Groups in Central Tennessee," manuscript in preparation for Tennessee Division of Geology.

is in Wells Creek basin in southeastern Stewart County, which has been described by Bucher;[6] the other is located along Flynn Creek about 5 miles south of Gainesboro, Jackson County, recently described by Wilson and Born.[7] (See Fig. 1.)

STRUCTURAL FEATURES OF THE HOWELL DISTURBANCE

The structural features of the Howell area may best be described in three parts: (1) underlying, intensely deformed rocks, including Black River and Trenton strata; (2) Fernvale formation; and (3) Chattanooga shale and Fort Payne chert. The first series is separated from the second by a marked nonconformity with maximum differential relief of approximately 100 feet within $\frac{1}{2}$ mile (Fig. 3). The "plane" of this nonconformity coincides with the pre-Fernvale surface. The Fernvale (and Brassfield) formation appears to have essentially disconformable relations with the overlying Chattanooga shale and Fort Payne chert.

UNDERLYING, INTENSELY DEFORMED STRATA

The much brecciated rocks of Black River and Trenton age are limited to a circular area about 1 mile in diameter. As the strata of these groups occur in blocks varying in size from small fragments up to 20 feet or more in dimension and abutting each other at all variants of strike and dip, details of a structural pattern were not determined. Many of the individual blocks show contortion and warping of bedding planes. In the creek just north of Howell, blocks of the Hermitage formation have been rotated in respect to each other with resulting small-scale thrust-faulting. Such faulting, however, is restricted to contacts between adjacent blocks and is too closely spaced and irregular to be mapped.

These blocks of limestone are imbedded in a matrix of shatter breccia composed of smaller fragments of limestone in a groundmass of powdered limestone. The breccia and the powdered limestone have been forced to flow around the blocks and along fractures within them, somewhat as in dike intrusion.

[6] W. H. Bucher, "Cryptovolcanic Structures in the United States," *XVI Int. Geol. Cong. Rept.* (1933), pp. 1066–70.

[7] C. W. Wilson, Jr., and K. E. Born, "The Flynn Creek Disturbance, Jackson County, Tennessee," *Jour. Geol.*, Vol. XLIV (1936), pp. 815–35.

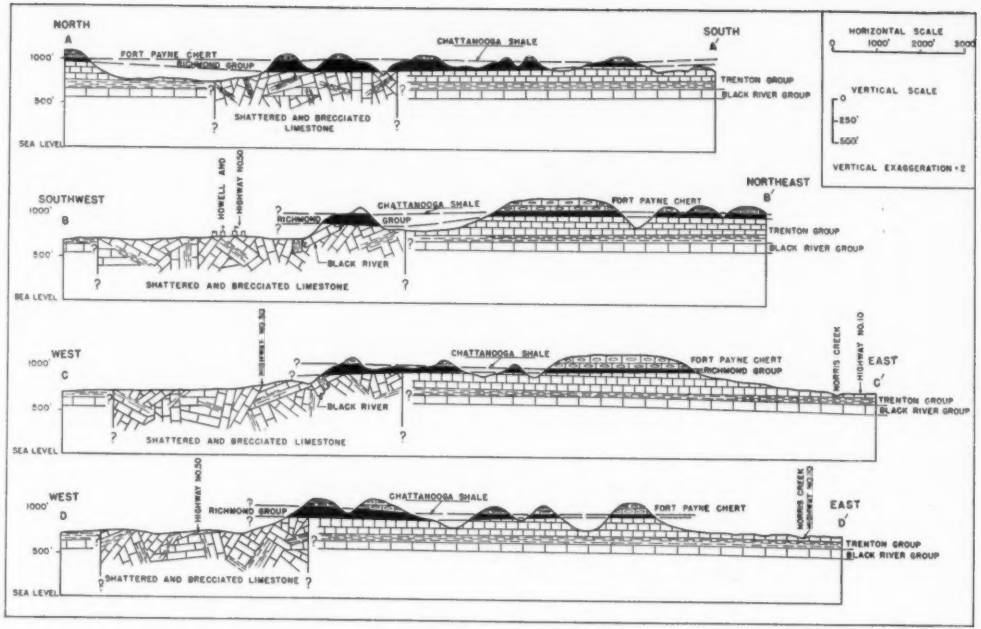

FIG. 3.—Structural cross sections of the Howell structure, along lines *A–A'*, *B–B'*, *C–C'*, and *D–D'* on Figure 2

This circular area of jumbled, brecciated, and unbrecciated limestone has been uplifted vertically in part, so that blocks of Carters limestone are now in juxtaposition with the surrounding undisturbed Cannon limestone outside the brecciated area. The maximum uplift was approximately 100 feet. Some blocks of Trenton limestone occur at the same level, or even below, their normal horizon; but these blocks are believed to have fallen or rolled to these positions at the time of origin of the crater.

A rather definite break occurs between the jumbled and brecciated limestone within the circumference of the Howell disturbance and the surrounding normal limestone of the Cannon and Catheys formations. Strikes and dips in the surrounding normal limestones were recorded, but these proved to be so irregular that generalizations as to whether they dip away from or in toward the disturbed area were not attempted.

FERNVALE FORMATION

The Fernvale formation rests nonconformably upon the underlying, greatly deformed Trenton and Black River strata. Unfortunately, Fernvale strata are preserved only over the eastern arc of the disturbed circular area and to the east of this area. The extensive removal of Fernvale prevents conclusive determination of its former extent, thickness, and structural details. However, contrasting the preservation of the Fernvale in the mapped area with its usual absence in the northern half of Lincoln County,[8] it is believed that this preservation is due to unusual local conditions. Also, the rapid thickening of Fernvale from 15 to 115 feet toward the deformed area suggests some genetic relationship between the local abnormal thickness of Fernvale and the deformed area. It is believed, therefore, that a crater, probably closed—although this cannot be confirmed—approximately 100 feet[9] in depth and 1 mile in diameter, existed here in pre-Fernvale times. This crater must have been exposed to appreciable erosion before Fernvale deposition began, for its sides were not steep but rather were graded, as indicated by the abnormally thick Fernvale extending southeastward beyond the circumference

[8] Wilson, *op. cit.*, Fig. 1, p. 450.

[9] This depth represents the depth immediately before the beginning of Fernvale deposition. The original depth—i.e., pre-erosion depth—was undoubtedly greater.

of intense deformation that probably marked the limits of the crater (Fig. 3). The graded, open crater was then flooded by the Fernvale sea, and filling with Fernvale sediments began. Deposition of the lower shale unit did not continue long for the sea was soon freed from silt, and the clear-water limestone unit was deposited. After a maximum of 35 feet of this limestone unit had been deposited, the sea began bringing in large quantities of silt and quartz pebbles from some distant source. These ultimately filled the crater; and, as the sea subsequently regressed, streams draining the recently emerged, muddy coastal plain cut channels and filled them with sand.

The brecciation of the Fernvale limestone unit has been referred to and reasons given for believing it to have been in part the result of contemporary brecciation. One significant fact is that, even though locally brecciated, this limestone is a continuous unit that was deposited over the strongly deformed rocks. Also, the elevation of the Fernvale limestone averages lower within the circular area of deformation than outside, where it overlies normal strata, indicating that the Fernvale did not participate in the uplift that locally raised rocks of the Black River and Trenton groups above their normal levels.

The thick shale unit that overlies the limestone unit locally has dips as high as 45°, and yet the top and bottom of this unit are traceable around the valleys and spurs with no more than expected variations in elevation. It is believed that these dips resulted partly from slumping and tilting within the shale, owing to readjustment and settling of blocks in the underlying deformed strata. However, it may be necessary to postulate a mild post-Fernvale and pre-Chattanooga renewal of activity to account for such high dips. If so, this would probably be renewal of the same forces that caused the pre-Fernvale crater, because the same area was affected.

CHATTANOOGA SHALE AND FORT PAYNE CHERT

These two formations show no brecciation. Although they are warped, having a structural relief of 60 feet on the base of the Chattanooga shale within the area mapped, the warping is very irregular and apparently has no relation to the pre-Fernvale crater.

AGE OF DEFORMATION

The localized forces which brecciated the Black River and Trenton limestones obviously operated after the deposition of the Catheys formation. As the Leipers formation is not present today in this mapped area, it is impossible to date the brecciation relative to this formation. However, the Leipers formation is believed to have covered most, or all, of central Tennessee; hence, a post-Catheys, pre-Leipers crater should have been filled with Leipers sediments. On this assumption the deformation would have been post-Leipers and would have been followed by removal of all Leipers in the vicinity by pre-Richmond erosion.

The Arnheim formation is thought to have been less extensive than the Leipers[10] and therefore may never have been deposited in what is now the northern part of Lincoln County. It is impossible, therefore, from any known data, to determine the age of the deformation with reference to the Arnheim formation.

Even though the Fernvale limestone unit is locally brecciated and high dips occur in the Fernvale shale unit, the major deformation, when viewed from a study of all known facts as well as these anomalies, would appear to have been pre-Fernvale. There is no basis for argument for a post-Fernvale date for the maximum deformation, but there is some basis for believing in a post-Fernvale renewal of the activity, which was so great in pre-Fernvale times. If this did occur, it would have an important bearing on the problem of origin of the deformative forces; but, unfortunately, the data are not sufficient to prove that the initial strong pre-Fernvale deformation was followed by a mild post-Fernvale renewal of deformation.

In summary, it is believed that the major deformation in the Howell disturbance may be dated as post-Catheys (probably post-Leipers) and pre-Fernvale, with possibly post-Fernvale and pre-Chattanooga recurrence in a mild form.

RESULTS OF MAGNETIC SURVEY OF THE HOWELL STRUCTURE

A magnetic survey of the Howell area was made by John C. McCampbell, of Vanderbilt University, with an Askania vertical magnetometer. The results are summarized in an isogammal map (Fig. 4).

[10] Wilson, *op. cit.*, p. 479.

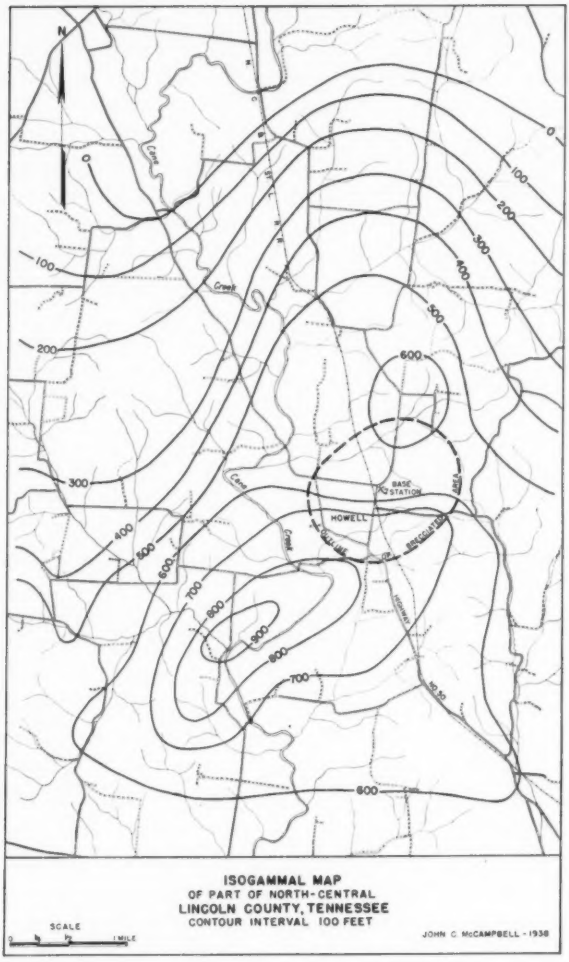

FIG. 4.—Magnetometer map of part of Lincoln County, Tennessee. The contours connect points of equal magnetic intensity. The position of the brecciated area is shown by the heavy broken line.

The center of a magnetic high with "closure" of 335 gammas was found $1\frac{3}{4}$ miles southwest of the center of the Howell structure. From the shape, size, and position of this high it cannot be shown definitely that it is related to the Howell disturbance, but it becomes more significant when compared with a similar offset to the southwest of the magnetic high associated with the Flynn Creek disturbance.[11] At Flynn Creek the offset is 4 miles to the southwest. The distances of the magnetic high from the center of disturbance in these two cases are proportional to the diameters of the respective disturbed areas, as the Flynn Creek disturbance has a diameter of 2 miles, compared with 1 mile for the Howell disturbance. Although this agreement of magnetic highs lying to the southwest of the two disturbed areas may be fortuitous, it is at least suggestive of buried igneous plugs, the presence of which would support a cryptovolcanic hypothesis of origin. However, it is possible that these magnetic highs are the reflections of buried meteoritic masses lying to the southwest of the craters produced by their striking the earth's surface obliquely from the northeast.

Origin

Any acceptable theory of origin for the structural features in the Howell area must explain the following: (1) a circular mass of jumbled and brecciated limestone, part of which has been uplifted approximately 100 feet relative to surrounding strata; (2) the shattering of the Black River and Trenton limestone into blocks and the irregular jostling of these blocks; (3) the pulverizing of much of the limestone into "rock flour"; (4) the unusual ability of breccia and rock powder to force their way into fractures; and (5) the formation of a crater 1 mile in diameter and more than 100 feet in depth, centered over the brecciated area.

These features characterize the group of structures described as cryptovolcanic by Bucher,[12] as well as the Wells Creek basin and the Flynn Creek disturbances in Tennessee. In 1936 the writers attributed the Flynn Creek disturbance to a subsurface volcanic explosion, in agreement with the explanation stated by Bucher, as follows:

[11] Wilson and Born, *op. cit.*, Fig. 8.

[12] *Op. cit.*, pp. 1055–74.

The structures here described must be considered unsuccessful attempts of volcanic materials to break through to the surface. Violent explosions are usually regarded as phenomena characteristic of the last stages of volcanism, blowing away the volcanic mountains that had piled up above the orifice in the early stages. But similar effects are to be expected where ascending columns of lava failed to break through to the surface. If in such circumstances the magmas are highly charged with gases, especially with water vapor, the gases may dissipate through joints and other fractures. Occasionally, however, such an escape may be prevented by the texture of the rocks, and the gases may accumulate near the surface. Then the pressure will rise to the point where it lifts the overlying rock columns. Mere lifting will allow the gas to escape, chiefly through fractures in the center, causing the characteristic disruption and jumbling of blocks and the upward movement recorded in cryptovolcanic structures. If enough pressure is accumulated, however, the overlying column is blown out, producing explosion funnels at the surface. . . . .[13]

Since publication of the description of the Flynn Creek disturbance Boon and Albritton[14] have pointed out the striking similarity between certain American "cryptovolcanic" structures and those which would be produced by impacts of giant meteorites. They conclude that explosive "backfiring," due to the elasticity of the rocks, would account for the unusual structural features associated with these disturbances as well as, if not better than, the cryptovolcanic hypothesis. In April, 1937, Boon and Albritton[15] enlarged upon geological results of meteorite impacts and refer to the Flynn Creek disturbance as a possible example of meteorite crater. They believe that their hypothesis is able to explain the salient features of this disturbance with the exception of the magnetic high located about 4 miles southwest of the disturbed area, whose location they think may be fortuitous.

Washburne[16] has stated that the main difference between cryptovolcanic structures and the salt-dome type of structure is the greater number of faults that can be mapped on many cryptovolcanic struc-

[13] *Ibid.*, pp. 1074–75.

[14] J. D. Boon and C. C. Albritton, Jr., "Meteorite Craters and Their Possible Relationship to 'Cryptovolcanic Structures,' " *Field and Laboratory*, Vol. V (1936), pp. 1–9.

[15] "Meteorite Scars in Ancient Rocks," *ibid.*, 1937, pp. 53–64.

[16] Chester W. Washburne, "Salt Domes, Meteor Craters, and Cryptovolcanic Structures," *Bull. Amer. Assoc. Petr. Geol.*, Vol. XXI (1937), pp. 629–30.

tures. He states that both types of faulted domes require local concentration of vertical stress but that, in his opinion, "neither requires the violent explosive action inferred by Bucher." The present writers do not agree with this conclusion, for they believe the Flynn Creek and Howell craters, with associated injected breccias and powdered limestone, require extremely violent explosive action. Neither of the present writers has seen Upheaval Dome, in southeastern Utah, referred to by Washburne; but they do believe that Jeptha Knob, Kentucky, Serpent Mound, Ohio, and Wells Creek basin, the Flynn Creek disturbance, and the Howell structure, Tennessee, were not formed as a result of salt-doming. In a region where salt beds are unknown, either at the surface or in subsurface drilling records, salt domes could hardly be expected to occur. Furthermore, salt-doming is not believed to be sufficiently explosive to blow out a crater 2 miles in diameter and 300 feet deep, as in the Flynn Creek disturbance.

Rust[17] published a very interesting paper on explosive volcanism in southeastern Missouri; and, although not referring to the cryptovolcanic hypothesis, he does present data that have a possible bearing on that conception. His account gives a picture of what could well have occurred beneath a cryptovolcanic structure and is strikingly in accord with this explanation. The debris-choked explosion tubes in Missouri, however, are on a smaller scale than any described cryptovolcanic structure, and also, owing to greater denudation, a much lower phase of "cryptovolcanism" is seen there than in these described structures in Ohio, Kentucky, and Tennessee.

CONCLUSIONS

The writers recognize difficulties in both the cryptovolcanic and the meteoritic hypotheses and, for the time being, prefer to maintain as neutral a position as possible until more data are found. Unfortunately, the Howell disturbance does not present new features that will aid greatly in determining the origin of this group of structures. The occurrence of a magnetic high $1\frac{3}{4}$ miles southwest of the Howell disturbance probably adds weight to the cryptovolcanic hypothesis. Also, the strong possibility of post-Fernvale and pre-

[17] George W. Rust, "Preliminary Notes on Explosive Volcanism in Southeastern Missouri," *Jour. Geol.*, Vol. XLV (1937), pp. 48–75.

Chattanooga renewal of the same localized force that formed the pre-Fernvale crater would support the cryptovolcanic hypothesis.

Summary of Events in the Howell Area

1. Deposition of Carters, Hermitage, Bigby, Cannon, and Catheys formations. If the Leipers formation were deposited prior to the explosion, as is believed to have been the case, it was removed by postexplosion erosion. No evidence is available for dating the explosion with regard to the Arnheim formation.
2. An explosion, blowing out a crater at least 100 feet in depth and 1 mile in diameter, and piling up limestone debris around the crater.
3. Removal of this debris (and possibly the Leipers formation from the surrounding area) and the grading of the crater walls by erosion.
4. Deposition of the Fernvale formation, filling the crater level with the surrounding floor of the Fernvale sea.
5. Possibly mild renewal of the force that produced the pre-Fernvale explosion.
6. Erosion interval.
7. Deposition of Brassfield limestone.
8. Erosion, leaving only a few local remnants of Brassfield chert.
9. Deposition of the Chattanooga shale and Fort Payne chert.

Acknowledgments.—This study was sponsored by the Tennessee Division of Geology, and the writers offer their thanks to Walter F. Pond, State Geologist. Mr. Pond has read the manuscript and offered useful criticism. Professor R. T. Chamberlin kindly offered suggestions that greatly improved the manuscript, and for this the writers wish to acknowledge their indebtedness. The writers also wish to thank John C. McCampbell for furnishing the results of his magnetic survey of the Howell area. John Webb and Herman Ferguson assisted in the field work.

THE FAUNA OF THE LYSOROPHUS POCKETS IN THE CLEAR FORK PERMIAN BAYLOR COUNTY, TEXAS

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ABSTRACT

Nodules of firmly cemented red clay containing remains of *Lysorophus* and associated forms have been obtained from eight localities in the Clear Fork formation of Texas. A new amphibian, *Euryodus primus* gen. and sp. n. is described in detail. It is referred to the order Lepospondyli and is shown to have nectridian affinities. Evidence suggesting a considerable size range in the amphibian *Lysorophus* is presented. The fauna of the "lysorophus pockets" consists of several types of small lepospondylous amphibians. A lungfish tooth also has been found associated, but life-association is doubtful. The fauna probably lived in rain pools along the beds of intermittent streams. It is more Pennsylvanian than Permian in aspect.

INTRODUCTION

During the summer of 1938 a field party from Walker Museum, in charge of the writer, collected a large number of nodules containing remains of *Lysorophus*. Altogether, seven localities were visited, all in the Clear Fork formation in Baylor County, Texas. A study of this material, as well as that collected by the writer and others in previous years, has yielded some interesting information concerning the fauna associated with *Lysorophus*. It is the purpose of this paper to describe a new amphibian found in these deposits, to discuss forms heretofore unknown in the nodules, to indicate the various localities in the area from which *Lysorophus* has been obtained, and to summarize the faunal assemblage of these pockets.

ORDER LEPOSPONDYLI

SUBORDER NECTRIDIA(?)

Euryodus N. GEN.

Genotype.—*Euryodus primus* n. sp. Walker Museum No. 1565. Skull and lower jaws.

Diagnosis.—Skull completely roofed over except for large orbits and nares and very small otic notch. No pineal foramen. Postorbitals and postfrontals very long and narrow. Tabular large. Supra-

temporal and intertemporal absent. Interparietal large and not incorporated into occipital plate.

Parasphenoid long and broad, tapering anteriorly. Interpterygoidal vacuity small. Palate well roofed by bone.

Upper dentition consisting of simple, conical teeth, with one, the seventh in genotype, greatly enlarged.

Lower jaws heavy.

Euryodus primus N. SP.

Holotype.—Same as genotype.

Horizon and locality.—Clear Fork formation, lower Permian. West bank of Brushy Creek, Baylor County, Texas.

Diagnosis.—Skull length, 33 mm.; skull width (temporal), 23 mm.; interorbital width, 11 mm. Thirteen teeth in upper jaw. Teeth 1–6, inclusive, are simple cones of relatively uniform size; average length 1.2 mm., average width at base 0.5 mm. Seventh tooth greatly enlarged; length 3 mm., width 2 mm. Teeth 8–13 successively smaller, with eighth about the same size as sixth.

Paratypes.—The following forms found in associated nodules are designated as paratypes. Walker Museum No. 1566: skull with 7 vertebrae articulated; shows additional details of occiput and temporal region. Walker Museum No. 1567: 2 skulls, one with good palate. Walker Museum No. 1569: partial lower jaw.

Description.—The following description is based on the holotype and the paratypes.

Dorsal surface of the skull: The pattern of the skull is clearly shown in Figure 1, *A* and *B*, and needs no further description.

Dentition: The condition of the upper dentition for the holotype has been given in the diagnosis. The paratypes do not vary from this pattern. The lower dentition consists of a series, number undetermined, of anterior teeth of uniform size, a large tooth, and 3 very small posterior teeth.

Occiput: The occipital plate is rather narrow. It lies essentially vertically and is pierced by a large, triangular foramen magnum. The elements are not separated by sutures. The exoccipitals appear to be the elements which comprise the double occipital condyle, but the basioccipital is well ossified.

Palate: The details of the structure of this region are not very clearly shown in any of the specimens; however, the general outlines

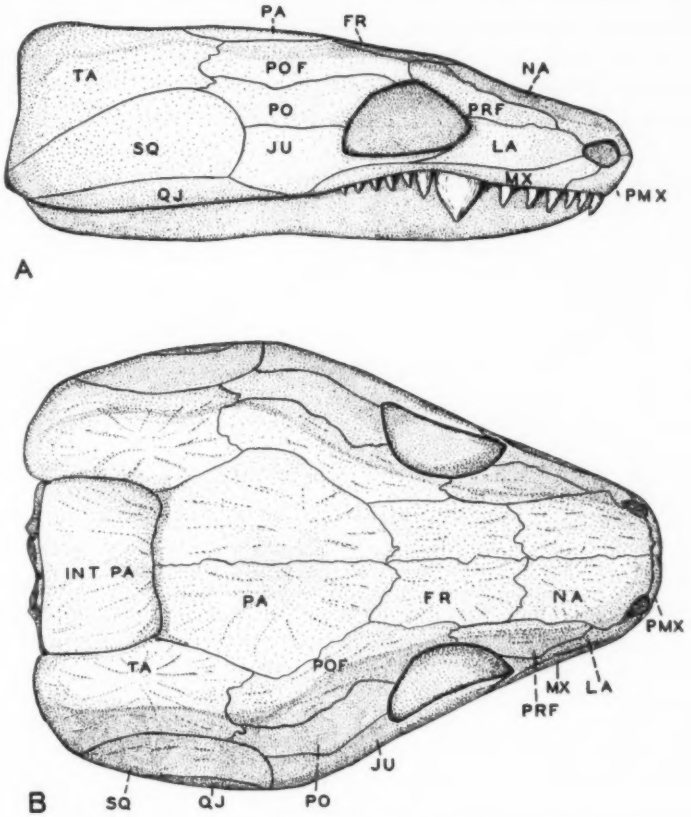


FIG. 1.—Restorations of the skull of *Euryodus primus; A*, dorsal; *B*, lateral (×3, approx.). Abbreviations: *FR*, frontal; *INT PA*, interparietal; *JU*, jugal; *LA*, lacrimal; *MX*, maxillary; *NA*, nasal; *PA*, parietal; *PMX*, premaxillary; *PO*, postorbital; *POF*, post frontal; *PRF*, prefrontal; *QJ*, quadratojugal; *SQ*, squamosal; *TA*, tabular.

can be seen in No. 1567. A reconstruction of the palate from this specimen, with a few details added from No. 1566, is shown in Figure 2. Noteworthy characters are the small size of the interpterygoidal

vacuity, the lack of firm connection between the basipterygoid process and the pterygoid, and the completeness of the palatal roof. The only palatine teeth observed lie near the lateral margins of the palate in a single row a short distance back of the internal nares.

Lower jaws: The lower jaws are quite heavy and pass from the snout to the very back of the skull. The point of articulation with the skull indicates that the quadrate occupied a position under the

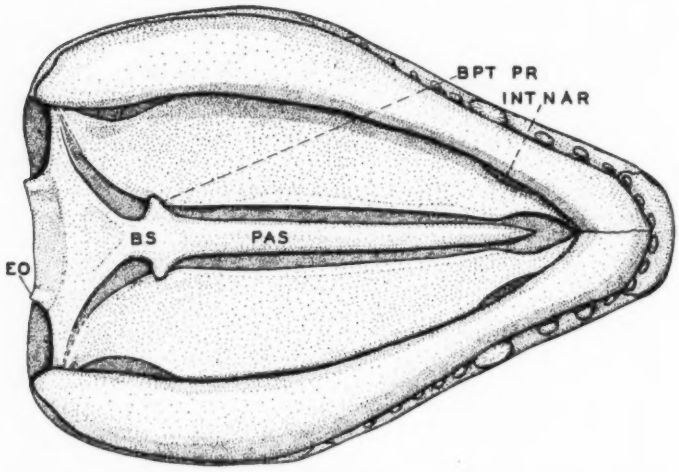


FIG. 2.—Restoration of palate of *Euryodus primus* (×3, approx.). Abbreviations: *BPT PR*, basipterygoid process; *BS*, basisphenoid; *EO*, exoccipital; *INT NAR*, internal nares; *PAS*, parasphenoid.

squamosal at the back of the skull. The bone itself is not visible in any specimen. It has not been possible to work out with any certainty the sutures between the various bones of the lower jaws.

Vertebrae: The vertebrae articulated with the skull of specimen No. 1566 are best seen in ventral aspect. There is a single central ossification similar to that in such lepospondyls as *Diplocaulus* and *Crossotelos*. This element is similar in size and proportion to that in *Lysorophus* but does not have any suggestion of the ventral grooving so evident on the centra of the latter. The atlas of *E. primus* is short and flares anteriorly to form two broad facets which articulate with

the double condyle of the skull. It appears to be similar in shape to the corresponding element in *Diplocaulus*. The remaining vertebrae show little differentiation. By sectioning one of them, it has been determined that the central ossification is notochordal and that the neural spine was single and not divided as in *Lysorophus*. There are two transverse processes for the articulation of the ribs, one on the arch and one on the centrum. The dorsal process is strongly developed.

Relationships.—Most of the small amphibians which have been found in the Clear Fork seem to be fairly closely related to *Lysorophus*. *Platyops* and *Diplocaulus*(?) *pusillus* are exceptions, both being close to *Diplocaulus*. Both the vertebrae and the skull immediately exclude *Euryodus* from any close relationship with *Lysorophus*. Characters of the vertebrae have been compared. The skull differs most markedly in the temporal and cheek regions, which are relatively free of bone in *Lysorophus* and completely covered over, except for the otic notch and orbit, in *Euryodus*. Further, the articulation of the lower jaws differs radically, being far posterior in *Euryodus* and shifted anteriorly in *Lysorophus*. There seems no possibility that these two genera are to be included in the same suborder.

There are certain points of similarity between the vertebrae of *Euryodus* and *Diplocaulus*—sufficient, perhaps, to suggest that *Euryodus* may be tentatively referred to the suborder Nectridia. The atlas, as noted, is quite similarly shaped in the two forms. The central elements differ but slightly except for size. The neural spines of *Euryodus* are somewhat elongated, suggesting vaguely nectridian affinities. No caudal vertebrae are known, so that the condition of the hemal arches cannot be determined.

The skull of *Euryodus* differs markedly from that of *Diplocaulus* and excludes the former from a place in the family Diplocaulidae—in fact, from any recognized family of the suborder Nectridia. The relatively primitive condition of the palate is suggestive of a general nectridian affinity but does not place the form more definitely. Until further evidence is available, it appears best to refer *Euryodus* tentatively to the suborder Nectridia but to make no family assignment.

FAUNAL ASSOCIATIONS OF THE "LYSOROPHUS POCKETS"

Euryodus primus occurs in direct association with *Lysorophus*. It is at present known from one pocket near the head of Brushy Creek, but certain small vertebrae found elsewhere suggest the possibility of a more widespread occurrence. By a study of various pockets it has been possible to find evidence of a rather extensive series of forms which lived associated with the very abundant amphibian *Lysorophus*. The fauna studied in the preparation of this paper comes from eight localities scattered over the areas drained by Brushy Creek, East Coffee Creek, and West Coffee Creek. The localities are as follows:

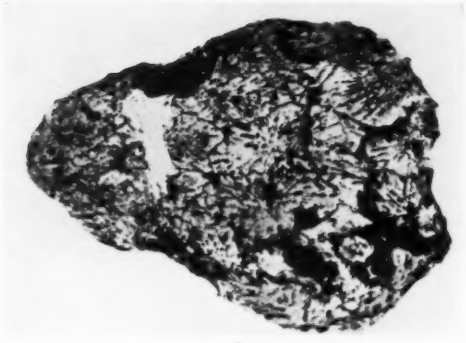
East Coffee Creek:
- Loc. No. 1: west banks, 1½ miles north of Lake Kemp
- Loc. No. 2: head of west branch of creek

West Coffee Creek:
- Loc. No. 1: head of creek
- Loc. No. 2: head of creek, ½ mile west of location No. 1
- Loc. No. 3: "Lysorophus hill," east banks of creek

Brushy Creek:
- Loc. No. 1: west bank, ¼ mile north of head of stream
- Loc. No. 2: west bank, ¾ mile north of location No. 1
- Loc. No. 3: Craddock's Ranch

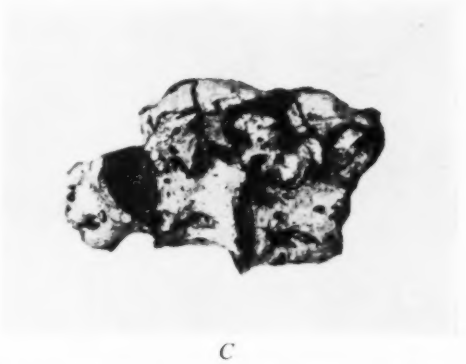
In nodules from all these localities, ribs and vertebrae of the typical small *Lysorophus tricarinatus* are abundant. In addition, a considerably larger type of vertebra is constantly present, sometimes common and sometimes rare (Pl. I, *C*). Often these vertebrae are associated in series of four or five, and in one instance there has been found a column over 6 inches in length. The vertebrae and associated ribs are much like those of *L. tricarinatus*, and the mode of preservation is similar. These vertebrae have generally been neglected in a study of the deposits in question and have usually been referred to *Diplocaulus*, if noted at all. A study of them by means of sectioning and from some very well-preserved specimens shows that they are very similar to the vertebrae of the small *Lysorophus*. They are entirely similar in the division of the neural arch and spine into lateral halves and in the deep, ventrolateral grooving of the centrum. As in *Lysorophus* the arch is free from the centrum, the only difference being that in the specimens studied the arch of the larger

PLATE I



A

B



C

A, Skull of holotype of *Euryodus primus* in dorsal aspect. ×2.

B, Skull of holotype of *E. primus* in lateral aspect. ×2.

C, Vertebrae of *Lysorophus* (?) sp. in lateral aspect. Specimen showing separation of neural arch and central element and pitting of centrum. Locality, Craddock Ranch.

form was less well attached than that in the smaller. The only actual difference that has been noted is that of size, the larger vertebrae running from about 5 to 9 mm. in length, much larger than any referred to *L. tricarinatus*.

In a nodule from East Coffee locality No. 2 a skull of *Lysorophus* somewhat larger than any which had come to my attention previously was found. Its total length is about 33 mm., the exact length being uncertain because of weathering of the snout. As far as can be determined, this form is identical, except in size, with *L. tricarinatus* and may tentatively be referred to the species as *L.* cf. *tricarinatus*. It indicates that *Lysorophus* does reach a considerable size but does not give evidence of any form large enough to accompany the vertebrae in question. With this indication of the size which may be attained by *Lysorophus* in mind, there is some justification for considering that the large vertebrae may also belong to the genus or to one closely related. Until further information on the characters of the form is available, it seems best to consider these large vertebrae as representing *Lysorophus*(?) sp.

In addition to these two common amphibian types, other forms are known from the "lysorophus pockets," all rather rare. Case[1] notes *Gymnarthrus* and *Cardiocephalus*(?) as small amphibians found associated with *Lysorophus*. They are rather similar to *Lysorophus* and are probably rather closely related. *Diplocaulus*(?) *pusillus* and *Euryodus primus* are other known associates, referred, with the reservations previously cited in the case of the latter, to the suborder Nectridia. Both are rare, although vertebrae may possibly be quite common but unrecognized. In all probability, *Platyops parvus* was associated with *Lysorophus*, although the data are not sufficient to make this certain.

Occasional scraps of larger amphibians have been found in some of the nodules, although they are quite rare and always badly worn.

From the Brushy Creek locality No. 2 a single, well-preserved lungfish tooth, *Gnathorhiza serrata*, was obtained by the writer in 1936. The association cannot be doubted; but there is, of course, the possibility that the tooth was washed into the deposits before solid-

[1] E. C. Case, "The Permo-Carboniferous Red Beds of North America and Their Vertebrate Fauna," *Carnegie Inst. Wash. Pub. 207* (1915).

ification of the clay and that the dipnoian did not actually live under the same conditions as did *Lysorophus* and the associated amphibians. The presence of but a single tooth suggests that this may be the case. The tooth is of interest in itself in representing the only known remains of dipnoian from the Clear Fork formation.

Summarizing the fauna known from pockets which are characterized by the presence of *Lysorophus*, the following forms are found to be present:

Fish:
Gnathorhiza serrata—rare
Amphibians:
Lysorophus tricarinatus—very abundant
L. cf. *tricarinatus*—1 skull
L.(?) sp.—abundant
Gymnarthrus willoughbyi—rare
Cardiocephalus sternbergi—rare
Euryodus primus—common in one locality
Diplocaulus(?) *pusillus*—rare
Platyops parvus—rare, association not certain
Fragments of indeterminate larger amphibians
Reptiles:
None unless some of the larger scraps may be referred to this class

Life-association of all the forms, with the possible exception of the lungfish and the probable exception of the larger amphibians, seems highly probable. The nodules in which the bones are preserved consist of red clay, cemented in all except one case by calcite. Silica forms the cement in nodules from locality No. 2 on West Coffee Creek. Mud, very similar to that which must have formed the nodules, is being deposited today in rain pools along the beds of the intermittent streams in the same area. The red color is derived, of course, from the surrounding red beds. The source of the color in Permian times is uncertain, and therein may lie some difference between the formation of the muds of that time and those of today. Otherwise the conditions of deposition of the "lysorophus pockets" may well have been very similar to those now active. The restricted nature of the individual deposits suggests some such a mode of deposition. The fauna of small amphibians and perhaps small fish is

comparable in general nature to that found in some of the larger pools today. As has been suggested by previous writers, intermittent desiccation of the pools may account for the great abundance of *Lysorophus*, a form probably unable to leave the water, and the relative scarceness of other small amphibians, able to seek new areas during particularly dry spells.

No other fauna closely comparable to the one described above has been found in Permian deposits of North America. The typical labyrinthodonts of the Permian are replaced in these deposits by lepospondyls. In general constitution the fauna is somewhat more like various faunas of the Pennsylvanian than like a typical Permian assemblage.

THE PLACE OF THE IOWAN DRIFT

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ABSTRACT

There appears to have been a concerted effort on the part of certain glacial students and their followers to give in textbooks and other publications an air of finality to the interpretation that the Iowan drift is an early Wisconsin product and to practically exclude the alternative interpretation, held by certain other glacial students, that it is a late product of the Illinoian stage of glaciation. The present paper aims to present for consideration evidence which seems to the writer to favor the view that the Iowan represents a dying-out phase of glaciation rather than an opening phase, and also that at the place where the Iowan and Illinoian drift come together, in southeastern Minnesota, there appears to be no evidence of a difference in age shown by weathering or by overriding of the Illinoian drift by the Iowan.

The Iowan drift is a product of the Keewatin or central Canada part of the North American ice sheet. Its gathering ground embraces about the same area as that of the late Wisconsin drift of Iowa and Minnesota, and it has only a limited exposure outside that drift and chiefly adjacent to the Des Moines lobe of Wisconsin drift. But none of its students has referred it to the late Wisconsin substage. It has been referred either to an early or opening part of the Wisconsin glacial stage or to a late part of the Illinoian stage in the recent classifications. In earlier classifications it was made a distinct glacial stage between the Illinoian and the Wisconsin stage. It is admitted by all students to be distinctly older than the late Wisconsin drift.

In general, the Iowan drift is a remarkably meager deposit, as now preserved, and is held by most students, if not all, to have been originally a thin deposit, making a scanty veneer on the greatly eroded surface of the Kansan drift. The veneer is so thin that the drainage lines in the Iowan area are practically coincident with those of the pre-Iowan drainage. In this respect it is in striking contrast with the drainage on the late Wisconsin drift, for that drift so completely filled the old valleys that new drainage lines had to be opened on it after the ice sheet disappeared.

The Iowan has no distinct recessional moraines, and its border is so indefinite on the east side that none of the students has been able to fix its limits. On the west border, however, it has sufficient def-

initeness to be located within a space of a mile or two and in places has low ridges and closely grouped knolls of morainic type. The writer thus found little difficulty in mapping this border for a distance of 200 miles, from Florence, South Dakota, where it emerges from beneath the late Wisconsin drift, to the place near Odebolt, Iowa, where it passes under the late Wisconsin. The course and character of the border are shown in Figures 7, 8, and 9 of *Professional Paper No. 161, United States Geological Survey.*[1]

While the Iowan drift is generally a thin deposit, its area embraces several strips of gravel knolls that lie generally in the valleys that had been cut in the Kansan drift prior to the Iowan glaciation. They run toward the border of the Iowan drift and not parallel with it, as in the case of recessional moraines. These knolls, as a rule, exhibit far less leaching of limestone pebbles in their surface portion than is found to characterize gravel deposits on the surface of the Kansan drift, and there seems to be a general consensus of opinion that they should be included in the Iowan deposits. They are seldom so sharp and linear as typical eskers. They form what may be termed a kame belt. In the writer's opinion they pertain to drainage within a relatively stagnant ice sheet. In my latest report on this region I made the following statement:

> The Iowan drift is so scanty and its limits in northeastern Iowa, as well as in southeastern Minnesota, are so ill-defined, that its extent is still an open question. There seems also to be an entire absence of recessional moraines. These conditions strongly suggest a stagnation or nearly complete cessation of movement in the Iowan ice sheet soon after it had reached its culminating position. The belts of gravelly ridges, which generally have a trend toward the border of the drift instead of parallel with it, are consistent with and seem to support the interpretation of a stagnation of ice movement.[2]

On page 35 of the same report the following statement appears in reference to the part of the Iowan drift exposed west of the Des Moines lobe of late Wisconsin drift:

> Although the Iowan drift in this district west of the Des Moines lobe of Wisconsin drift has a more definite border than that east of this lobe, it seems

[1] Frank Leverett and Frederick W. Sardeson, "Quaternary Geology of Minnesota and Parts of Adjacent States," *U.S. Geol. Surv. Prof. Paper No. 161* (Washington, D.C., 1932).

[2] *Ibid.*, p. 28.

to have no recessional moraines and is generally a very scanty deposit. These conditions make it seem probable that this part of the Iowan ice sheet also became stagnant soon after reaching its culminating position.

For the consideration of glacial students I am now presenting the view that a condition of stagnation in an ice sheet is more natural for a waning or dying-out phase of a glacial stage than would be natural for the opening phase of a glacial stage. If the Iowan is a late phase of the Illinoian glacial stage, as I have maintained in my latest report, there seems nothing surprising to have it reach a stagnant condition. But, if it comes as the opening phase of the Wisconsin glacial stage, it seems to me unnatural for it to become stagnant. If we consider the remarkable activity of the Illinois glacial lobe of early Wisconsin time, the nearly contemporary action in a neighboring (Iowan) lobe would seem to require activity rather than stagnation. In the Illinois lobe of early Wisconsin time the outer or Shelbyville moraine has a general relief of nearly 100 feet above the level of the outlying Illinoian drift in Illinois and continues prominent eastward into Ohio. This lobe was also forming definite moraines in the course of a recession of the ice border that preceded a readvance to the position reached at the building of the prominent Bloomington morainic system, the Cerro Gordo and Champaign morainic belts having been formed in the course of that recession of the ice border. For many thousand years this lobe displayed remarkable activity, following, as well as during, the culminating stage.

It seems hardly necessary to indulge in speculations as to whether the place of the Iowan is at the opening of the Wisconsin until we have looked into the evidence as to age that the drift itself offers. Attention will be given first to the place where the Iowan and Illinoian drifts make their closest approach, which is in southeastern Minnesota near the upper end of Lake Pepin. In that place the upper drift on the north side of Cannon River is a very stony red drift brought in from the north, while that on the south side of the river is clayey gray drift from the Keewatin center brought in from the west. Both drifts are underlain by a much older clayey drift from the Keewatin center, which extends eastward into Wisconsin about to the west border of the Driftless Area. In western Wisconsin are

exposures of a soil and weathered zone under the stony red drift that were developed on the clayey older drift. This old clayey drift when traced westward into Minnesota is found to be the Kansan drift, for it is there separated from an underlying still older drift by a definite soil and weathered zone, the older drift being the Nebraskan. The stony red drift therefore is post-Kansan and is the third drift of the series displayed in that district. It has been referred to the Illinoian stage by the writer because it seems to have experienced a degree of weathering similar to that exhibited by the Illinoian drift, as widely displayed in Illinois and adjacent parts of Wisconsin and Iowa. The characteristics of the part in Minnesota are briefly discussed on pages 22–23 of *Professional Paper No. 161*. The Wisconsin part was discussed in considerable detail by Dr. Samuel Weidman in his "Geology of North Central Wisconsin," published in 1907.[3] Dr. Weidman found a prominent moraine, termed the "Marshfield moraine," in Marathon, Wood, and Clark counties, which he mapped for a distance of 70 miles. He interpreted it to be "one of the older sheets of the drift series." He found only one drift sheet outside of it and so labeled it the "Second Drift Formation." But, when traced westward into Minnesota, as already indicated, it is found to be the third drift of the series and placed in the Illinoian glacial stage.

It lies like a blanket on the eroded surface of the Kansan drift and covers the valley bottoms and slopes as well as the uplands. In amount it is generally so slight that the streams on it are flowing in essentially the same courses that they had opened in the Kansan drift prior to its deposition. The average thickness in Dakota County, Minnesota, and in the neighboring part of Wisconsin is scarcely 10 feet. But in Dakota County, Minnesota, it forms a prominent range of hills that runs from Hampton southeast nearly to Cannon River. The hills are 75–100 feet high and form an intricate network a mile or more in width. The name "Hampton moraine" has been applied to it.[4]

[3] *Wis. Geol. Surv. Bull. 16*, pp. 451–56, Pls. 46, 47, and 48.

[4] It is of interest to note that Dr. R. T. Chamberlin, working as my assistant in 1906, was the discoverer of this moraine and at the time seemed much impressed with the weathered condition of its drift compared with that displayed by the neighboring moraines of Wisconsin drift.

The topmost drift south of Cannon River also veneers a greatly eroded Kansan drift as a thin sheet. It stands as the third drift of the Keewatin center of glaciation and is a continuation of the Iowan drift of northeastern Iowa. Its features from Iowa northward to this locality are discussed in some detail on pages 23–29 of *Professional Paper No. 161* and were given much careful thought in their presentation. A strip of knolly drift on it in northwestern Wabasha County is discussed on page 26 and is the nearest approach to terminal moraine found anywhere on this eastern border. The degree of weathering is much less than in the Kansan drift and consistent with that generally found on the Iowan drift. This great eastward extension of the Iowan drift at the high altitude of 1,160–75 feet, 500 feet above the adjacent Mississippi Valley, led Dr. Weidman and the writer to investigate carefully as to evidence of an overriding of the Hampton Illinoian moraine by Iowan ice. But we were unable to find any indication of such an overriding. The Illinoian lobe north of Cannon River seems to have held the ground to the exclusion of the Iowan lobe from the Keewatin center, thus making the one an exact correlative of the other.

This district is one that the writer has urged several geologists who have favored putting the Iowan in the Wisconsin glacial stage to give a careful personal investigation, and I here repeat the suggestion. It was the habit of the late Professor Calvin to claim that Iowa held the key for unlocking the whole Pleistocene history, but the writer, though born in Iowa and naturally interested in upholding its reputation, believes that the matter under discussion can be better worked out in Minnesota and Wisconsin than in Iowa; also that the way in which the Wisconsin ice sheet, and also the Illinoian, had a westward growth in its late stages is shown in Minnesota but not in Iowa. The group of glacial students who have been referring the Iowan to an early part of the Wisconsin seem to have adopted Calvin's view and to have shown remarkably little interest in districts outside Iowa and Illinois and southern Wisconsin. As to the classification of the loess deposits, eastern Nebraska seems to have a better key than Iowa, as I have indicated in a review of Dr. Lugn's *Bulletin No. 10, Nebraska Geological Survey*, "Pleistocene

Geology of Nebraska" in the *Journal of Geology* the past year.[5] The conditions governing the development of gumbotil and contrasts in the rate and degree of weathering of a given drift sheet require wider study than is afforded in the area with which these geologists are familiar. The gumbotil, as suggested by Marbut, is well developed chiefly in the district where shale forms a notable constituent of the drift and is not a conspicuous feature where sandstone and limestone constituents dominate. It also is undeveloped on Iowan drift within the shale area because of the topographic conditions, a flat surface, needed for its retention, being wanting.

The development of a pebbly concentrate on the Iowan drift has been given more significance by the writer than by this group of geologists, it being their contention that very little time would be needed to develop it. This is a matter that seems to require a more careful study than has thus far been given it. I have been very much impressed with the rapid rate at which slope wash has been going on in Iowa since the cover of tall prairie grass has been removed and the bare surface exposed to washing by heavy rains. Several inches of black soil now are to be seen as a coating on the pale soil in the valley bottoms. This feature greatly impressed Professor Calvin, and he frequently called attention of visiting geologists to it. If the present, bare slopes were to be taken as characteristic of the time when the pebble concentrate was forming, I am ready to grant that but little time might be required in its development. But, if a heavy coating of prairie grass got a foothold very soon after the ice sheet melted, the development may have been very slow. In the greater part of the Iowan area this pebble concentrate has a silt cover from several inches to several feet in thickness. This is an indication that the rate of reduction by slope wash was less rapid than the rate of deposition of dust. This dust seems to be a more recent deposit than the main part of the loess in outlying districts.

This seems a fitting place to call attention to the surprising differences in the degree of weathering exhibited by the Illinoian drift. In central Illinois disintegration of the pebbles for several feet from the surface involves the granite as well as the limestone pebbles, but in western Pennsylvania the granite pebbles in the Illinoian drift,

[5] Vol. XLVI (1938), pp. 784–86.

and even within 5 feet of the surface, still retain a surface polish and show very little evidence of disintegration. This is the more surprising in view of the very great erosion that has been effected by the Allegheny River, a deepening of the valley fully 200 feet in rock since the Illinoian drift was deposited. This amount of deepening is also found on the Ohio about down to Moundsville, West Virginia, beyond which it decreases gradually to a relatively small amount before reaching the mouth of the Big Kanawha River. The condition of the granite pebbles and boulders on or near the surface of the Iowan drift is markedly fresher than on the Illinoian drift of central and southern Illinois and somewhat fresher than near the surface of the Illinoian drift of southeastern Iowa. The latter has a much thicker cover of loess than the Iowan drift and also has a definite gumbotil, 1–3 feet thick, under the loess. Granite pebbles are scarce down to this depth. To what extent the better-drained condition of the Illinoian drift in western Pennsylvania, and also of the Iowan drift, is responsible for the lesser decay of pebbles seems a matter worth investigating.

This also seems a fitting place to give attention to a paper by Dean G. F. Kay that deals with the place of the Iowan drift and estimates the time involved in each of the several glacial and interglacial stages.[6] While Kay makes reference to the alternative view that the Iowan is a product of the Illinoian stage of glaciation, his sections given in Figure 1 put the Loveland loess *under* the Iowan but *above* the Illinoian, in keeping with a paper he had recently published in the *American Journal of Science*.[7] On pages 444–45 he speaks of the need for a revision of the Pleistocene classification and concludes that the time is ripe for it. The revised classification which he presents and illustrates by map and diagram on pages 449–52 puts the Wisconsin and Iowan and the intervening Peorian loess into a single cycle called the Eldoran epoch and refers the Sangamon interglacial and the Illinoian glacial stages to an earlier cycle named the Centralian. This does not admit for consideration the unsettled

[6] "Classification and Duration of the Pleistocene Period," *Bull. Geol. Soc. Amer.*, Vol. XLII (1931), pp. 425–66.

[7] "The Relative Ages of the Iowan and Wisconsin Drift Sheets," *Amer. Jour. Sci.*, Vol. XXI (5th ser., 1931), pp. 158–72.

question as to the place for the Iowan. His classification has the further fault of adding to the complexity of the nomenclature by introducing the new names applied to each of the epochs. On page 456 he asserts that the Iowan drift is not old enough to have gumbotil formed on it and does not make allowance for the unfavorable topographic conditions for its retention. On page 460, in estimating post-Iowan time from the depth of leaching, no recognition of removal of leached material in the development of the pebbly concentrate on its surface is given. So the post-Iowan time may be much longer than the 55,000 years which he estimates.

The estimates of the duration of glacial stages presented in this paper make them surprisingly brief—3,000 years each for the Wisconsin and Iowan, 9,000 years for the Illinoian, and 7,500 years each for the Kansan and Nebraskan. The time estimates given appear to be restricted to the time the ice sheets were in Iowa. While the late Wisconsin and the Iowan ice lobes terminated within the state, the Kansan and Nebraskan lobes extended entirely across Iowa into Missouri, and adjacent parts of Nebraska, Kansas, and Illinois, and thus embraced a considerable part of the glacial stage.

Instead of 3,000 years, the entire Wisconsin stage may have covered a period of 70,000 years or more, and it seems likely that the Illinoian, with Iowan included, embraced a similar period and that a similar length should be given the Kansan and Nebraskan glaciations. If this was the case, about 280,000 of the 700,000 years which Kay estimates for the length of the Pleistocene period, or 40 per cent of its length, was under glacial conditions. It seems probable that a period of 10,000–15,000 years would be required for the Wisconsin ice sheet to have reached its culmination at the Shelbyville moraine, and that 35,000–40,000 years would be involved down to the late Wisconsin culmination in the Port Huron morainic system, which dates back about 25,000 years, 15,000 of which will have been taken in the final retreat of the ice. Sardeson's estimate, from a study of the recession of St. Anthony Falls, that Lake Agassiz began flowing to Hudson Bay more than 8,000 years ago gives a rough indication of the date to be assigned to the close of the Wisconsin glacial stage.[8]

[8] Leverett and Sardeson, *op. cit.*, pp. 145–46; also *U.S. Geol. Surv. Minneapolis–St. Paul Folio* (Field ed.; Washington, D.C., 1916), pp. 86–88.

I recently had occasion to realize the unfortunate effect this time estimate by Kay is having when I read the thesis prepared by a graduate student of Washington University for the degree of Doctor of Philosophy. When I inquired why he had made use of Kay's time chart, he replied that, while he realized that the time given was far too short, Kay's chart seemed to be the only one available, and so he took it.

In closing, a few words seem appropriate on the efforts of certain glacial students to give finality to the interpretation that the Iowan

TABLE 1

Wisconsin glacial	Mankato drift (late Wisconsin)
	Cary drift (middle Wisconsin)
	Tazewell loess
	Tazewell drift (early Wisconsin)
	Iowan loess
	Iowan drift (possibly Illinoian)
Sangamon interglacial	
Illinoian glacial (perhaps including the Iowan drift)	
Yarmouth interglacial	
Kansan glacial	
Aftonian interglacial	
Nebraskan glacial	

drift is an early Wisconsin product and to practically exclude the alternative interpretation referring it to the Illinoian glacial stage. This is particularly unfortunate since it may be reproduced in textbooks and books intended for the general reader.

I will cite two instances in which a slight change in tables and names would overcome this objectionable feature. In a recent paper by MacClintock and Richards,[9] the statement, "*the* current classification of glacial deposits," etc., might better read "*a* current classification," etc. If then in Table 1 the names in parentheses, added by the writer, were used, this slight addition might make a presentation of the alternative classification unnecessary. The names, "Tazewell," "Cary," and "Mankato" seem an unnecessary addition to an

[9] Paul MacClintock and H. G. Richards, "Correlation of Late Pleistocene Moraine and Glacial Deposits of New Jersey and New York," *Bull. Geol. Soc. Amer.*, Vol. XLVII (1936), p. 291.

already complex nomenclature and less significant than the terms "early," "middle," and "late" Wisconsin already in general use.

The *Outline of Glacial Geology* by F. T. Thwaites,[10] though making reference to the alternative interpretation, has adopted elsewhere the classification of the Iowan in the Wisconsin and called it the "First Wisconsin." The Shelbyville moraine is then labeled "Second Wisconsin." It would seem wise to have avoided the labels "First," "Second," etc., so that, if the Iowan is later found to fall in the Illinoian instead of the Wisconsin stage, the Shelbyville and later morainic systems of the Wisconsin will not need to be renumbered.

Since writing this paper, I have received the March, 1939, number of the *Bulletin of the Geological Society of America*, containing Dean Kay's paper "Pleistocene History and Early Man in America." In the diagram (Fig. 2) showing Pleistocene stages, I note that the mere removal of the diagonal partition between the Iowan and the Illinoian will make them appear to be correlative deposits of the third stage of glaciation, that being the rank of each of them in the glacial series.

In this latest paper Kay repeats his estimate that the combined length of the glacial stages is 30,000 years. This, of course, cannot be harmonized with my estimate of 280,000 years. In this paper Kay apparently makes no question of the finality of his interpretation that the place of the Iowan is in the Wisconsin glacial stage. No reference is made to the alternative view of its Illinoian age—a view still standing in the literature. He appears to have been rather hasty in considering it an antiquated viewpoint.

[10] Ann Arbor, Mich.: Edwards Bros., 1934.

LOCH BOROLAN LACCOLITH, NORTHWEST SCOTLAND

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ABSTRACT

The scattered literature relating to this eruptive mass has been brought together and condensed into a connected account. The main part of the mass has the form of a laccolith which appears to rest on a floor of limestone. The silica content of the rocks decreases progressively from the top to the base of the laccolith, and the rocks near the base develop melanite, diopside, and nepheline. The variety known as "borolanite" contains also what appear to be pseudomorphs after leucite. Fundamental problems of petrogenesis are involved in the interpretation of the evidence.

INTRODUCTION

The complex of alkaline rocks at Loch Borolan, in the northwest Highlands of Scotland, raises some of the most interesting problems of petrology. Among them are (*a*) density stratification in an intrusive sheet; (*b*) desilication of acid magma by assimilation of limestone; (*c*) the genesis of nepheline; and (*d*) the nature and significance of "pseudoleucite." The literature relating to this complex is quite extensive, but it is scattered through many publications, some of which are not readily accessible.

Early notices by Heddle, Calloway, and Bonney may be dismissed as having only historic interest. A short paper by J. J. H. Teall[1] in 1886 contains what is perhaps the earliest suggestion that a magma may change its composition by assimilating limestone.

In 1892 there appeared a joint publication by J. Horne and J. J. H. Teall, "On Borolanite, an Igneous Rock Intrusive in the Cambrian Limestone of Assynt, Sutherlandshire."[2] This gives a detailed description of the most remarkable rock in the area, the so-called borolanite, and makes the suggestion that the white spots are pseudoleucite. Geological relations were not fully discussed, but it was shown that borolanite is associated with the syenite of Cnoc-na-Sroine and that "a zone of crystalline marble can be traced for long distances in immediate contact with or close to the eruptive

[1] "Note on Some Hornblende-bearing Rocks from Inchnadampf," *Geol. Mag.*, 1886, p. 346.

[2] *Trans. Roy. Soc. Edinburgh*, Vol. XXXVII (1892), p. 163.

rock." A further paper by Teall[3] in 1900 recorded the discovery of a rock containing fresh nepheline at the north foot of Cnoc-na-Sroine and gave additional support to the pseudoleucite theory.

In 1906 the present writer,[4] then a student in Germany, published a description of a large collection of rocks from the Borolan area.

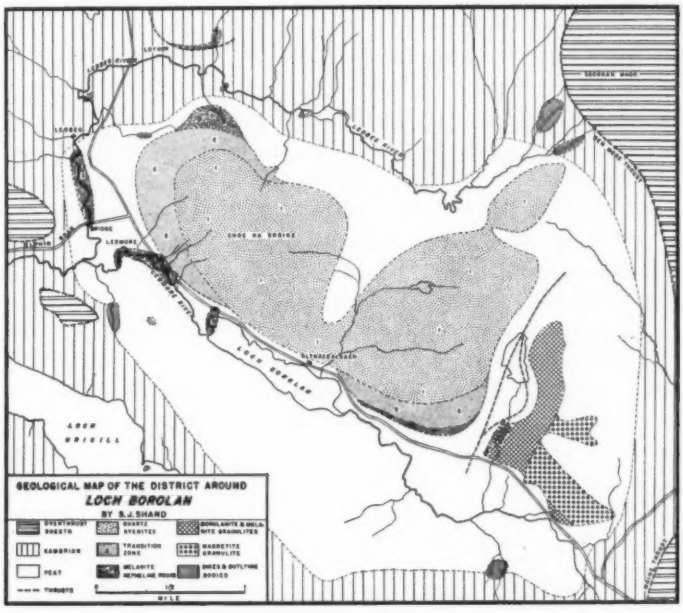

FIG. 1

Most of these had been collected by Professor Karl Busz, but a few additional specimens were supplied by Dr. Teall. Of the relations between the various rocks in the field the writer had at that time no firsthand knowledge; but during the next four years he made several visits to the locality, mapped it on a large scale, and made extensive collections for further study.

[3] "On Nepheline-Syenite and Its Associates, etc.," *Geol. Mag.*, 1900, p. 386.

[4] S. J. Shand, "Ueber Borolanit, etc.," *Neues Jahrb. Min.*, Beilage Band XXII (1906), p. 413.

In 1907 the great Memoir of the Geological Survey of Great Britain on the *Geological Structure of the Northwest Highlands of Scotland*, made its appearance. The section dealing with petrology, written by J. J. H. Teall, included an account of the Borolan area based on the publications described above.

The writer,[5] in 1908, published a preliminary account of his re-examination of borolanite and the pseudoleucite question; he followed it up in 1910 with a general account of the geology and petrology of the laccolith,[6] omitting much of the detail already given in the 1906 and 1908 papers. A number of chemical analyses, which were made for the writer by A. Gemmell,[7] were published as a separate article in the same volume. An independent analysis of borolanite by W. Campbell Smith[8] was published in 1909.

No further major contribution to the petrology of the Borolan area has appeared since 1910, but J. Phemister's account of the Loch Ailsh area,[9] published in 1926, has an important bearing on the matter since the two areas are only a few miles apart and yield similar rocks. The same writer, in a note published in 1931,[10] showed that a certain small outcrop of limestone with silicates, near the north end of Loch Borolan, is actually a carbonated eruptive rock.

STRUCTURE OF THE COMPLEX

The main body of eruptive rock forms a hill 4 miles long and 1–1½ miles wide; the highest part is called Cnoc-na-Sroine and rises to 1,306 feet. The hill is completely surrounded by a belt of low-lying peaty ground in which few exposures can be found. Beyond this belt, which is only 100 yards wide at the north end of the hill but as much as a mile wide on the west flank, the encircling rocks are dolomitic limestones and quartzites of Cambrian age. The "im-

[5] Shand, "On Borolanite and Its Associates in Assynt," *Trans. Edinburgh Geol. Soc.*, Vol. IX, Part III (1909), p. 202.

[6] "On Borolanite and Its Associates in Assynt (Second Communication)," *ibid.*, Part V (1910), p. 376.

[7] "Chemical Analyses of Borolanite and Related Rocks," *Trans. Edinburgh Geol. Soc.*, Vol. IX, Part V (1910), p. 417.

[8] "On the Composition of Borolanite, etc.," *Geol. Mag.*, 1909, p. 152.

[9] *Geology of Strath Oykell* (Mem. Geol. Surv. Gt. Brit., 1926).

[10] *Summary of Progress* (Geol. Surv. Gt. Brit., 1930), p. 58.

bricate structure" developed by the great overthrusts has obscured the relation of the sediments to the eruptive rocks, but in general it is the limestone or marble that lies nearest to the eruptive core. Dikes of borolanite and syenite cut the limestone at several points.

The hill itself is formed, in the upper part, of quartz-syenite which passes down into quartz-free syenite. This is succeeded at the foot of the hill—where the exposures are admittedly poor—by syenite with accessory melanite and then by dark-colored syenitic rocks containing much melanite and diopside ("aegirine-augite") as well as some altered nepheline. Regarding the passage from one zone to the next, Teall wrote in 1900 that "the evidence available suggests that the quartz-syenites shade into quartzless syenites and these again into nepheline-syenites." The writer agrees with Teall and thinks that the observed changes of texture and composition may be due to increasing nearness to the edge and floor of the intrusive body. But A. Harker has formed a different opinion, and in 1916 he wrote that "there is not a gradual transition but a sharp boundary between the several rock types"[11] at Loch Borolan.

In view of this difference of opinion it is useful to observe that at Loch Ailsh, according to Phemister, "the quartz-bearing and quartz-free rocks are not well differentiated from one another," and again that "the passage from shonkinite to pulaskite is very rapid, but so far as the writer could determine it is continuous."

The thickness of the eruptive mass is at least 1,000 feet in the middle, and it may be more. It must have thinned out rather rapidly toward the sides, and this suggests that the original form was that of a laccolith. The writer's conception of the shape and internal structure of the mass is shown in Figure 2. A reversed fault or thrust-plane cuts across the southeastern end of the hill and has brought up a great mass of the spotted borolanite and melanite-granulites. The basal zone of melanite-pyroxenite, shown in the cross sections, is a deduction from the observed increase of the dark minerals downward and from the appearance of a body of melanite-pyroxenite on the western margin of the peat belt.

[11] "Differentiation in Intercrustal Magma Basins," *Jour. Geol.*, Vol. XXIV (1916), p. 554.

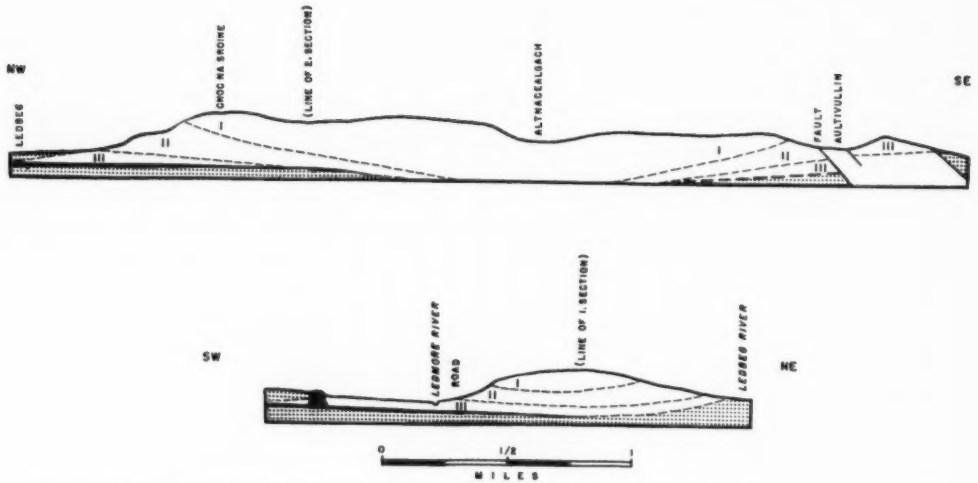

FIG. 2.—Two sections through the laccolith of Loch Borolan, Scotland. *I*, subacid granite, average content of lime 1–1.5 per cent; *II*, syenite, average content of lime 2–3 per cent; *III*, diopside, melanite, and nepheline rocks, average content of lime 8–14 per cent; *IV* (black), melanite pyroxenite, content of lime 22 per cent; *V* (dotted), Cambrian limestone.

PETROGRAPHY OF THE LACCOLITH

Zone 1.—The quartz-syenite is a highly leucocratic rock composed of pink or reddish feldspar, with abundant blebs of milky quartz in the most acid facies. As the proportion of quartz decreases, it becomes interstitial; the rock is then comparable with nordmarkite and holds some 10 or 12 per cent of quartz. The feldspar is partly microperthite and partly albite. Dark minerals are exceedingly scanty; they comprise scraps of bluish-green hornblende, a few flakes of biotite, and very little chlorite, epidote, iron oxides, zircon, and sphene. An analysis of the nordmarkite variety is given in the table of analyses (Table 1, col. 1).

TABLE 1*

CHEMICAL ANALYSES OF THE BOROLAN ROCKS

	1	2	3	4	5	6	7	8	9	10	11	12
SiO_2	68.00	62.96	45.06	33.52	56.26	44.85	48.19	43.22	33.49	70.46	53.13	52.85
Al_2O_3	16.49	21.26	20.95	7.60	21.93	17.10	18.52	15.14	9.10	17.01	24.87	16.55
Fe_2O_3	2.70	2.48	6.23	10.69	0.67	10.06	4.51	9.57	17.49	1.62	2.25	4.12
FeO	.....	0.46	2.84	6.98	.....	2.71	1.68	2.62	2.66	0.98	1.40	2.22
MgO	.....	0.19	3.31	6.94	.....	0.39	1.13	3.33	1.80	.....	.....	3.01
CaO	1.32	2.81	8.32	22.82	1.46	14.00	10.29	14.35	28.86	0.82	2.68	6.20
Na_2O	7.48	5.47	3.51	3.52	4.95	1.67	3.44	2.94	0.46	7.17	5.56	4.27
K_2O	3.45	3.49	4.09	1.37	10.63	6.92	8.05	7.18		1.92	7.26	4.25
H_2O	0.76	1.10	4.28	.....	4.16	1.66	3.45	.....	0.67	0.34	1.41	1.77
TiO_2	.....	0.33	1.25	4.51	.....	†	1.75	2.72	6.29	.....	0.60	2.30
P_2O_5	.....	.....	0.66	2.31	.....	.....	.....	.....	.....	.....	.....	1.07
Cl	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	0.77
Total...	100.20	100.55	100.50	100.26	100.06	99.36	101.00	101.07	100.82	100.32	99.16	99.38

* Too few of the analyses include P_2O_5 and Cl, and MnO was not determined in any instance.

† In analysis No. 6 titania was not separated from alumina. At least 2–3 per cent of titania must have been present in this rock.

1. Quartz-syenite (nordmarkite), Cnoc-na-Sroine. Analyst: S. J. Shand.
2. Melanite-syenite, north end of Cnoc-na-Sroine. Analyst: A. Gemmell.
3. Ledmorite. Analyst: A. Gemmell.
4. Cromaltite. Analyst: A. Gemmell.
5. White spot in borolanite, Aultivullin. Analyst: S. J. Shand.
6. Dark base of borolanite, Aultivullin. Analyst: S. J. Shand.
7. Typical spotted borolanite, Aultivullin. Analyst: W. Campbell Smith.
8. Borolanite-granulite, Aultivullin. Analyst: A. Gemmell.
9. Melanite from pegmatite, Ledmore. Analyst: A. Gemmell.
10. Aegirine-aplite, Ledmore. Analyst: A. Gemmell.
11. Foyaite, north end of Cnoc-na-Sroine. Analyst: A. Gemmell.
12. Assyntite, north end of Cnoc-an-Sroine. Analyst: A. Gemmell.

Zone 2.—The quartz-syenite passes downward into a zone of coarse and generally rather rotten red or gray syenite composed al-

most entirely of microperthite and alteration products. Sometimes a little interstitial quartz is present, sometimes a few grains of partly decomposed melanite and biotite; and sometimes a little sodalite makes its appearance in the feldspar. An analysis of melanite-syenite from the north end of Cnoc-na-Sroine is given in Table 1, column 2.

Zone 3.—This is a zone of melanite-pyroxene syenite with subordinate nepheline, which is poorly exposed about the base of the hill, especially in the banks of the Ledmore and Ledbeg rivers. The passage from Zone 2 to Zone 3 can be followed on the hillside, $\frac{1}{2}$ mile along the road beyond the north end of Loch Borolan. Close search by the writer failed to reveal any positive evidence of discontinuity in this section. There is a rather rapid diminution in grain-size, which may be a consequence of the thinning-out of the laccolith at its edges, and a quick increase in color index; but neither of these changes is so sudden as to prove discontinuity.

The rock exposed in the banks of the Ledmore River is an even-grained aggregate of orthoclase, melanite, pyroxene, and micaceous pseudomorphs after nepheline, the average grain-size being 1 mm. The orthoclase is filled with curious "fingerprint" structures formed by threads of a micaceous mineral running through the feldspar. The threads are rounded in cross section and have a diameter between 0.01 and 0.005 mm. The original mineral in the threads is thought to have been sodalite. Melanite may form as much as 25 per cent of the rock; it is brown to yellow in color and is full of tiny crystals and granules of sphene which have been formed by transformation of melanite. In varieties that carry little melanite, the latter has crystallized very late and forms allotriomorphic masses surrounding and enclosing pyroxene and even orthoclase. The pyroxene is acmite-diopside (so-called aegirine-augite). Some green biotite is present too.

The rock is certainly a curious and unusual one, and the writer, following the fashion of the time, felt justified in giving it a new name; so he called it *ledmorite*. A chemical analysis of this rock appears in column 3 of Table 1.

The rock seen in the Ledbeg River is similar except that pyroxene predominates and melanite is quite scarce.

Zone 4.—This hypothetical basement zone is a deduction from the observed downward concentration of the dark minerals, and more specifically from the appearance, on the edge of the marble belt west of the Ledmore River, of a body of melanite-pyroxenite. This is a coarse-grained, dark-green rock composed of pyroxene, melanite, and biotite similar to those in ledmorite, with some ilmenite and apatite in addition. The three silicates are intergrown with one another, but the pyroxene shows the greatest tendency to be idiomorphic and the melanite least. The density of this rock is about 3.4. A chemical analysis by Gemmell (col. 4, Table 1) shows about 22 per cent of lime, a figure which greatly exceeds the highest amount ever recorded in a normal eruptive rock. To this unique rock the writer gave the name of *cromaltite*. Within this rock there are streaks and lenses of a gray, flinty material composed of calcium and iron carbonates mixed with silicates, ilmenite, and apatite. It was interpreted in 1910 as a "half-fused sediment which has absorbed a certain proportion of silicates from the intruded cromaltite," but J. Phemister has since demonstrated that the flinty material is in reality a product of replacement of silicates by carbonates and therefore without value as evidence of assimilation.

THE SOUTHEASTERN END OF THE LACCOLITH

Overthrusting from the southeast has brought up melanite-rich rocks, such as characterize Zone 3, to the level of the quartz-syenite in the main laccolith and has caused shearing, granulation, and mylonization. The principal rocks to be seen in this part are the spotted borolanite, with rounded and sometimes polygonal spots of "pseudoleucite"; sheared borolanite, in which the spots have been pressed out to white streaks; and granulitic borolanite or melanite-granulite, in which no trace remains of the white spots. In the final stage of mylonization the dark minerals have been completely altered to chlorite and magnetite, and the feldspar is considerably sericitized. There are also spotted rocks of the same general appearance as the spotted borolanite but which contain no melanite at all; these may represent borolanite that has suffered hydrothermal alteration without crushing.

A rock of entirely different character which intervenes between

the borolanite area and the quartz-syenite area to the northwest has the features of a hornstone. It contains rounded grains of diopside and poikilitic plates of hornblende and epidote in a fine-grained mosaic of orthoclase and oligoclase. This rock has undoubtedly been thrust up into its present position, and, if it is a silicified limestone, as some of its characters suggest, then it lends support to the view that the laccolith has a limestone floor.

Typical borolanite is a dark-gray rock with white spots—the supposed pseudoleucite—which stand out on weathered surfaces as nearly spherical bodies about ½ inch in diameter. Most of the spots have rounded or irregular outlines, but a minority of them, especially in the melanite-free rocks mentioned above, show traces of two or three facets and suggest the shape of leucite crystals. Under the microscope the spots are seen to consist of allotriomorphic grains of orthoclase and oligoclase, with rosettes of secondary mica and interstitial zeolitic matter. No characteristic internal structure is exhibited. A chemical analysis of one of these white spots is given in column 5, Table 1. The groundmass of the rock is a granular aggregate of orthoclase and melanite, with subordinate green biotite and a good deal of fibrous zeolitic matter of the composition of natrolite or mesolite.

The melanite, which makes up about one-third of the rock (by weight), is seldom perfectly idiomorphic but forms irregular, much-cracked masses. When freshest, it is deep brown, but the color often changes to yellow, and then little crystals and grains of sphene are developed within the melanite. This is not a case of enclosure of early formed crystals within later, but an actual transformation of melanite into sphene.

Analyses of spotted borolanite and of borolanite granulite are given in Table 1, columns 6, 7, and 8, and an analysis of melanite in column 9.

DIKE ROCKS

A few of these are worthy of brief mention. They fall into an older and a young group, as follows:

Older Group	Younger Group
Ledmorite-pegmatite	Aegirine-aplites
Spotted and unspotted borolanites	Foyaites with fresh nepheline, including the so-called "assyntite"
Aegirine pegmatites	

The ledmorite-pegmatite is chiefly remarkable for containing melanite crystals up to 7 mm. in diameter. These furnished the material for the analysis in Table 1, column 9.

Spotted borolanite forms dikes, cutting marble at the source of the Ledbeg River and at two other points farther down that river, in each case within marble; also at a point in the peat belt west of Loch Borolan. In these dikes the borolanite shows the same general character as in the main area, but the spots are more often pink than white, and some of them show hexagonal pseudomorphs of mica (after nepheline) enclosed within the larger orthoclase crystals. In three instances a transition can be seen from spotted borolanite into feldspathic syenite. This point is important in connection with the origin of the supposed pseudoleucite.

Aegirine-pegmatite with pseudomorphs after nepheline occurs in a few small dikes. Aegirine-aplite or microgranite forms strong dikes at fully six localities and cuts both ledmorite and borolanite. An analysis of aegirine-aplite appears in Table 1, column 10.

Foyaite with fresh nepheline was first discovered by Teall at the extreme north foot of Cnoc-na-Sroine, where the hill approaches the river most closely. The rock is a typical foyaite except that the only dark minerals are melanite and biotite, about 3 per cent of each. An analysis is given in Table 1, column 11.

The name *assyntite* was given by the writer in 1910 to another foyaitic rock which had previously been called "augite-syenite" (Teall) and "augite-sodalite-syenite" (Shand). In addition to orthoclase, nepheline, sodalite, and a more or less sodic pyroxene, this rock is distinguished by the presence of an unusual quantity of sphene, which forms poikilitic plates and skeleton crystals enclosing all the other minerals of the rock. In this rock, too, the "finger-print" structure (already mentioned in connection with ledmorite) is particularly well developed and it is possible to show that the original mineral in this structure is sodalite. An analysis of assyntite appears in Table 1, column 12.

THEORETICAL ISSUES

We are now ready to discuss the problems indicated at the beginning of this study. We have seen that there is a difference of opinion as to whether the rocks of the laccolith belong to a single

intrusion or to a series of successive intrusions, Teall holding one view and Harker the other. It is not difficult, in the field, to distinguish three zones on the basis of texture and composition combined; but close inspection shows that each zone grades into the next and that the limits indicated by texture do not coincide exactly with the limits indicated by mineralogical composition. Zone 2 contains a little quartz in its upper part and a little melanite and sodalite in its lower part. An imaginary surface separating all the quartz-bearing rock from all the melanite- and feldspathoid-bearing rock would have as its outcrop a wavy line somewhere about the middle of Zone 2, and it is quite certain that there is no physical discontinuity in this position. For this reason the writer shares Teall's opinion that the laccolith is the product of a single intrusion of magma. But even if it were not so, the position would not be greatly changed, for we should still have to explain the variation of quartz within Zone 1, the change from quartz-syenite to melanite-syenite in Zone 2, and that from ledmorite to cromaltite in Zone 3; that is, instead of explaining the vertical differentiation of the laccolith as a whole, we should have to explain the vertical differentiation of each zone separately.

It is important to realize that the observed differentiation has at least two independent components, an upward increase of free silica and a downward increase of lime and magnesia. The sinking of early formed crystals of diopside and melanite might account for the second component but could by no means influence the distribution of free silica. Besides, we have seen that melanite is a mineral of quite late crystallization in some of these rocks.

N. L. Bowen[12] has contended that, if the original magma was very rich in the components of potash-feldspar, with only a small excess of silica, then leucite might crystallize from it at an early stage. He calculates that leucite crystals would be slightly denser than such a magma and would therefore sink through it and accumulate in Zone 3, where they would transform into pseudoleucite (orthoclase and nepheline). Zone 1, being thus deprived of leucite, would crystallize with an excess of free silica.

[12] *Evolution of Igneous Rocks* (Princeton, N.J.: Princeton University Press, 1928), p. 255.

Now if Zone 3 had been composed entirely of the spotted borolanite, this ingenious theory would find much support; but except at the southeastern end Zone 3 is formed of ledmorite, which contains no pseudoleucite and has more soda than potash (in molecular proportions). And if the original magma was a highly feldspathic one, as Bowen supposes, where did the abundant melanite and pyroxene come from? Even if we granted all that Bowen claims, the problem would be only half-solved.

From Arthur Holmes we have a very different solution. He says:

> The whole assemblage is one which can be readily accounted for in terms of alkali-ultrabasic magma and the products of its advance emanations on the invaded quartzites and other quartz-rich sediments. Alkalies, probably present in part as aluminates, travelled furthest, forming leucocratic quartz-syenite and nordmarkite. The downward increase of melanite and sphene testifies to the less extensive migration of iron, calcium and titanium. Lower down calcium and magnesium appear in increasing amounts as pyroxene, showing that these elements, whether carried as magma or as emanations, did not rise so far. The nearest approach to uncontaminated "magma" appears to be represented by cromaltite.[13]

Observe that Holmes recognizes the presence of "quartzites and other quartz-rich sediments" in the area but completely ignores the limestone. It means nothing to him that the complex is intrusive in the Durness limestone horizon; that the most lime-rich members of the complex are found nearest to the limestone; and that dikes of borolanite cut the limestone at several points. It means nothing to him either that every mineral in the cromaltite encloses every other, as in a metamorphic lime-silicate rock; or that this rock, which he supposes to have given off aluminous emanations, has a very low content of alumina in relation to bases. It seems that Holmes was so engrossed with his "flux of emanations" hypothesis that he just dismissed from his mind all evidence that did not lend itself to his argument.

In the writer's opinion there is only one process that can account completely and rationally for the observed differentiation in the Borolan laccolith, and that process is assimilation of limestone at the base of the laccolith. There may or may not have been con-

[13] "The Petrology of the Volcanic Area of Bufumbira," *Geol. Surv. Uganda, Mem. III*, Part II (1937), p. 100.

comitant assimilation of quartzite at the roof, but we have no evidence on this point. If the original magma was an oversaturated one, corresponding perhaps to the mean composition of Zone 1, then reaction with the underlying limestone would bring about a diminution of silica toward the base and an accumulation of calcium-magnesium silicates there. Cromaltite presents such an accumulation, only slightly diluted with magmatic alkalis and alumina. Ledmorite marks a further stage of dilution of the lime-silicates by magmatic aluminosilicates. Among the latter, some nepheline makes its appearance, and under the circumstances it does not seem possible to account for it except by the desilication of albite molecules. The lower part of Zone 2 also shows a slight deficiency of silica; but, apart from a little melanite, this zone is nearly free from lime minerals. It is possible to see the operation of gravity in this matter. Either gravity carried crystallized lime-silicates down from Zone 2 into Zone 3, or else gravity prevented dissolved lime-silicates from diffusing upward, while permitting silica to diffuse downward. It is vain to try to follow the process in detail, but the general picture presented by the assimilation theory corresponds in every respect with the evidence seen in the field. As far as that evidence goes, it is entirely favorable to the view that the differentiation of the laccolith was governed by assimilation of limestone.

Only the pseudoleucite theory remains to be discussed. The writer in his earlier work supported the idea that the white spots in borolanite are pseudomorphs after leucite. Afterward he observed in the field three instances of typical spotted borolanite passing, in the space of a few centimeters, into coarse feldspathic syenite with subordinate melanite and nepheline, in such a way as to suggest that the spots are just isolated and rounded crystals and fragments of feldspar and nepheline. This observation seemed to throw doubt on the pseudoleucite theory; but without further evidence it is impossible to reach any final decision in the matter. Arthur Holmes says that the spots "appear to be complex intergrowths of felsic minerals replacing quartzite fragments, the replacement being due to the introduction of alkali aluminates from the invading magma." This suggestion finds no support whatever in the field evidence.

SQUEEZE-UPS

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ABSTRACT

Extrusion of viscous lava through vents and cracks in a solidifying flow produces squeeze-ups, which may be divided into bulbous and linear types. They are probably formed after the crust has attained a considerable thickness. Many of them are hollow owing to drainage of lava from the central part of the squeeze-up after its crust has hardened.

INTRODUCTION

Colton[1] has used the term "squeeze-ups" for grooved tongues of lava formed when viscous lava was squeezed up through fissures in the hardened crust of a flow. The author[2] has used the term "grooved squeeze-ups" for somewhat similar features. Jaggar[3] and others have called the bulbous squeeze-ups described below "toes," and Brigham has used the term "pushes" for them.

In this article any small feature on the surface of a flow formed by the extrusion of viscous lava through an opening in the solidified crust will be called a "squeeze-up." These squeeze-ups may be divided into bulbous and linear types, although forms transitional between them are common.

BULBOUS SQUEEZE-UPS

Bulbous squeeze-ups are common features on the surface of the McCartys flow[4] in New Mexico. Near the end of the flow they are usually not more than 5 feet in diameter and 2 feet in height. Commonly they are much smaller. However, near the source of the flow not only are they much more numerous but they are also as much as 10–15 feet across. They are formed by the upwelling of lava through

[1] Harold S. Colton and Charles F. Park, Jr., "Anosma or 'Squeeze-Ups,'" *Science* (N.S.), Vol. LXXII (1930), p. 579.

[2] Robert L. Nichols, "Grooved Lava," *Jour. Geol.*, Vol. XLVI (1938), pp. 601–14.

[3] T. A. Jaggar, "Lava Stalactites, Stalagmites, Toes, and 'Squeeze-Ups,'" *Volcano Letter 345* (1931), pp. 1–2.

[4] Nichols, "Quaternary Geology of the San José Valley, New Mexico," *Geol. Soc. Amer. Proc. 1933*, abst., p. 453.

vents at places where cracks in the crust are deeper and wider than usual or where two or more cracks intersect. A good example is shown in Figure 1. The shape clearly indicates that the lava was rather viscous when extruded, for, had it been more fluid, it would have spread out as a flat sheet forming a thin flow-unit[5] on the

FIG. 1.—A bulbous squeeze-up on the McCartys flow

surface of the flow. This squeeze-up is also compound, having been formed by the extrusion of three separate gushes. The left-hand bulb was formed first; it was later partly covered by the central bulb, which in turn was in part covered by the right-hand bulb. No clue was obtained as to whether the two later bulbs were formed by the bursting of the crust of the first bulb or whether they broke out separately.

In some places the surface is dotted with well-scattered squeeze-ups; in other places they are so close together that they give to the surface of the flow a characteristic knobby appearance, as is well

[5] Nichols, "Flow-Units in Basalt," *Jour. Geol.*, Vol. XLIV (1936), pp. 617–30.

shown in Figure 2. Near the source of the flow the surface is billowy, owing to numerous large squeeze-ups.

LINEAR SQUEEZE-UPS

Linear squeeze-ups formed by the upwelling of lava along cracks in the crust may be as much as 20 feet long and are usually 4–5 inches high. The cracks through which they were squeezed may be

FIG. 2.—Numerous bulbous squeeze-ups, giving the surface of the flow a knobby appearance.

formed by a contraction of the crust in cooling, by its collapse due to removal of support, or by compressional forces acting on it. As shown in Figure 3, the lava extruded was not in great quantity and was probably relatively viscous.

TIME OF FORMATION

The squeeze-ups were probably formed after the crust had attained a considerable thickness, for it seems reasonable to suppose that, if the crust had been thin, more lava would have been extruded and presumably the lava would have been sufficiently fluid to devel-

op a flow-unit rather than a squeeze-up. In fact, gradations between squeeze-ups and small flow-units are common. Had the crust been very thin at the time of the formation of the squeeze-ups, one would expect it to have been basined by the weight of the extruded lava. In no case was this phenomenon observed.

FIG. 3.—A linear squeeze-up on the McCartys flow

HOLLOW SQUEEZE-UPS

Many of the bulbous and linear squeeze-ups are hollow—a condition apparently caused by drainage of lava from the central part of the squeeze-up after its crust had hardened. This drainage is probably due either to a reduction in the quantity of lava reaching the liquid thread from upstream while the flow is advancing downstream or to a more rapid withdrawal of lava from the tube caused by rapid progress of the flow downstream. Such drainage is possible only if the vent or crack which fed the squeeze-up remains open.

VALUE IN CROSS SECTION

Squeeze-ups, where recognizable in the cross section, would be useful in differentiating both successive superimposed flows and flow-units.

ACKNOWLEDGMENTS.—The author takes pleasure in acknowledging the field assistance of Mr. Robert S. Folsom. Thanks are due to Mr. Ralph S. Fellows for the photographs and to Professor Kirk Bryan, of Harvard University, for guidance in the preparation of the manuscript.

THE MEAN CHEMICAL COMPOSITION OF METEORITIC ACCRETION

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ABSTRACT

From a discussion of the relative number and mass of metallic and stony meteorite falls, the ratio of mass accretion is considered to lie between 2:1 and 1:4. Since no precise value for the ratio is apparent, the mean composition of meteoritic accretion is computed for a range of ratios.

In a recent issue of the *Journal of Geology* Nininger[1] has called attention to the uncertainties involved in obtaining a significant mean composition for the meteoritic accretion of the earth. The difficulties arise principally in determining the relative amounts of metallic and stony matter which fall, for we should be able to fix separately, within narrow limits, the mean composition of the two materials. For the stony material a systematic error may develop through the inclusion of finds, since the more acidic stones are seldom collected unless they are seen to fall.[2] In the mean compositions determined by Farrington and Merrill[3] for stones the finds constitute respectively only 13 and 19 per cent of the analyses; hence, systematic errors are probably small. Furthermore, the agreement between various determinations of the mean composition indicates the essential correctness of the figures (cf. tables in Merrill's articles). In the following discussion we shall accept as approximately correct the separate mean compositions of metallic and stony meteorites and shall devote our attention to determining the relative masses of these materials now being accumulated by the earth.

The mean composition of all meteoritic material is not readily

[1] H. H. Nininger, "The Composition of Meteorites," *Jour. Geol.*, Vol. XLVI (1938), pp. 889–91.

[2] G. P. Merrill, "The Percentage Number of Meteorite Falls and Finds Considered with Reference to Their Varying Basicity," *Proc. Nat. Acad. Sci.*, Vol. V (1919), p. 37.

[3] Merrill, "Researches on the Chemical and Mineralogical Composition of Meteorites," *ibid.*, Vol. I (1915), p. 429; "Composition and Structure of Meteorites," *U.S. Nat. Mus. Bull. 149* (1930), p. 17.

determined. Clark's method, as applied by Farrington,[4] is not correct, for it weights the metallic and stony material according to the number of analyses made. Farrington combined analyses of 318 iron and 125 stone samples, assigning a weight of 72 to the irons and 28 to the stones; as a result, his "average composition" has little significance. Merrill[5] used the ratio between the total known masses of metallic and stony meteorites: 93 per cent metallic, 2 per cent stony-iron, and 5 per cent stony. This method, as in Farrington's reduction, does not differentiate between falls and finds and has, as Merrill recognized, little justification.

Our interest lies principally in the relative proportions of metallic and stony materials in space. Since the matter shed into the atmosphere during a meteorite's air flight will eventually reach the earth's surface, no differences will develop between the space and terrestrial quantities of the various meteoritic materials. W. J. Fisher[6] estimated, on the basis of the relative masses collected from falls, that the mass of metallic material, including stony-iron, exceeded that of the stony by 2.5:1. Millman[7] found that 64 per cent of the photographed spectra of sporadic meteors were of iron, a fact which suggests that sizable metallic particles are twice as abundant as comparable stony particles. Although Millman's identifications of stony and metallic meteors are probably correct, they cannot be accepted without reservation until certain of the criteria used by him are checked in the laboratory.

All meteorites undergo a terrific ablation and disintegration during their passage through the earth's atmosphere. As a result the stones, composed of rather friable material, are usually broken into several pieces. At times the disruption is so complete that great showers of small stones fall, e.g., at Mocs (3,000 stones), at Pultusk (100,000 stones), at Homestead (100 stones), and at Holbrook

[4] O. C. Farrington, "Analyses of Stone Meteorites," *Field Mus. Pub. 151* (1911), p. 213.

[5] "The Present Condition of Knowledge on the Composition of Meteorites," *Proc. Amer. Phil. Soc.*, Vol. LXV (1925), p. 119.

[6] "On the Ratio of Iron to Stone in Meteorite Falls," *Harvard Observ. Bull. 845* (1927), p. 5.

[7] P. M. Millman, "An Analysis of Meteor Spectra: Second Paper," *Harvard Observ. Annals*, Vol. LXXXII, No. 7 (1935), pp. 149–77.

(14,000 stones). The largest single stony mass to be recovered *in situ* was the Long Island (Kansas) fall, 564 kilograms, which broke upon

TABLE 1

AVERAGE COMPOSITION OF ALL METEORITES

ELEMENTS	COMPOSITION		RATIO METALLIC TO STONY			
	Metallic (Per Cent)	Stony (Per Cent)	2:1 (Per Cent)	1:1 (Per Cent)	1:2 (Per Cent)	1:4 (Per Cent)
H						
He						
Li						
Be						
B						
C	0.04	0.16	0.08	0.10	0.12	0.14
N						
O		36.30	12.00	18.00	24.00	29.00
F						
Ne						
Na		0.30	0.10	0.15	0.20	0.24
Mg	0.03	14.30	4.60	7.10	9.50	11.00
Al		0.76	0.25	0.38	0.50	0.61
Si	0.76	18.00	6.00	9.00	12.00	14.40
P	0.15	0.10	0.13	0.12	0.12	0.11
S	1.70	1.89	1.80	1.80	1.80	1.90
Cl		0.03	0.01	0.01	0.02	0.02
A						
K		0.07	0.02	0.04	0.05	0.06
Ca	0.05	1.30	0.47	0.67	0.88	1.10
Sc						
Ti		0.10	0.03	0.05	0.07	0.08
Va						
Cr	0.03	0.14	0.07	0.09	0.10	0.12
Mn	0.03	0.18	0.08	0.10	0.13	0.15
Fe	88.30	25.60	67.00	57.00	47.00	38.00
Co	0.53	0.14	0.40	0.33	0.27	0.22
Ni	8.15	1.40	5.90	4.80	3.60	2.80
Cu	0.05	0.01	0.04	0.03	0.02	0.02
Zn						

hitting the earth; in contrast, the largest iron masses weigh tons. Nininger's recent work[8] on the Archie and Pasamonte falls demonstrates the great amount of material which may be scattered beneath the path of a falling stony meteorite. As a result of such disintegration a considerable fraction of the original material will not be

[8] "The Meteorite Fall of August 10, 1932, near Archie, Cass County, Missouri," *Pop. Astr.*, Vol. XLIV (1936), p. 93; and "The Pasamonte, New Mexico, Meteorite," *ibid.*, 1936, p. 331.

recovered; hence, the average mass of a stone fall suffers in comparison with that of an iron. This result is, however, compensated by the very process which produces it. A shower of many stones over a wide area is more likely to become known as a "fall" than is the descent of a single iron mass, with the result that the frequency of stone falls is

TABLE 2

AVERAGE COMPOSITION OF ALL METEORITES EXPRESSED AS OXIDES

OXIDES	RATIO METALLIC TO STONY			
	2:1 (Per Cent)	1:1 (Per Cent)	1:2 (Per Cent)	1:4 (Per Cent)
SiO_2	14.00	20.00	26.00	31.00
TiO_2	0.05	0.08	0.10	0.13
Al_2O_3	0.96	1.40	1.90	2.30
Fe_2O_3	0.31	0.46	0.62	0.73
Cr_2O_3	0.19	0.24	0.29	0.34
Fe	62.50	50.00	37.00	27.00
Ni	5.80	4.60	3.40	2.50
Co	0.37	0.31	0.24	0.19
FeO	4.50	7.80	9.10	11.00
NiO	0.13	0.20	0.27	0.32
CoO	0.02	0.03	0.04	0.05
CaO	0.67	0.94	1.20	1.50
MgO	7.90	12.00	16.00	19.00
MnO	0.10	0.13	0.16	0.19
Na_2O	0.27	0.42	0.55	0.66
K_2O	0.05	0.08	0.10	0.13
P_2O_5	0.33	0.33	0.34	0.34
S	1.80	1.80	1.80	1.90
C	0.08	0.10	0.13	0.14
Cu	0.04	0.03	0.02	0.02

increased relative to that of iron falls. After a single stone is observed to fall, a search usually results in the discovery of many similar masses, which raise the total mass of the fall. These conflicting factors suggest that for stone falls the average mass will be low, while the total mass may be high, in comparison to similar quantities for iron falls. For 392 stone falls the total mass is 7,705 kilograms and for 21 iron falls 730 kilograms, a ratio of 10.5:1 for the total masses, or 1:1.8 for the average masses. Thus, it would appear that the ratio between the quantities of falling stony and metallic materials is probably between 1:1 and 4:1. The range of values taken for

this ratio in the past is illustrated by Farrington's use of 1:2.5, whereas V. M. Goldschmidt[9] has recently used approximately 4:1.

Inasmuch as at present no evidence allows a definite choice of the proportion of stony to metallic materials, the mean composition of all incoming meteoritic material has been computed for a range of ratios (Tables 1 and 2). Many elements present only as traces in meteorites have been omitted as being relatively unimportant. The data of Merrill[10] were taken for the stones, while the data of the Noddacks,[11] which are more complete than those of Farrington, were taken for the irons, after being corrected for 5 per cent troilite, FeS, removed prior to their analysis. The considerable range in composition, exhibited by the tables, may be of interest to geophysicists.

[9] "Die Mengenverhältnisse der Elemente und der Atom-Arten," *Skr. Norske. Viden.-Akad. Oslo, Math.-Nat. Kl.*, No. 4 (1937).

[10] "Composition and Structure of Meteorites," *op. cit.*

[11] I. and W. Noddack, "Die Häufigkeit der chemischen Elemente," *Naturwiss.*, Vol. XVIII (1930), p. 757.

DISCUSSION: WIND-DEPOSITION SHORELINES

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In his new classification of shorelines recently presented and defended, Shepard[1] proposes the term "wind-deposition" shoreline for a shore prograded into the sea by sand brought to place by wind. The present paper does not consider the merits of the classification nor does it attempt to explore all the possibilities with regard to "wind-deposition" shorelines. Obviously, in a locality where strong winds, blowing across an ample source of sand, dump sand into a water body having relatively ineffective waves and currents, prograding of the shore will take place. However, the writers view with skepticism the possibility that wind-blown sand has ever prograded a shore any great distance into the open ocean. It is, however, the purpose of this note not to discuss this general question but to redescribe and analyze the features of the Castle Neck area of northern Massachusetts cited by Shepard[2] as an example of this type of shoreline. Lucke[3] in his review takes issue on Castle Neck as a properly selected example.

The area is shown in Figure 1, redrawn from parts of the United States Geological Survey topographic sheets of Salem and Gloucester, rather than from the United States Coast and Geodetic Survey chart reproduced in Shepard's figure. Castle Neck is a stretch of sand nearly 3 miles long and more than ½ mile wide, extending E.SE., with a beach on the ocean side. It is covered with dunes and borders a great area of marsh on the south side. Figure 2 is an aerial photograph taken from the south which shows the Ipswich

[1] Francis P. Shepard, "Revised Classification of Marine Shorelines," *Jour. Geol.*, Vol. XLV (1937), pp. 602–24; John B. Lucke, "Marine Shorelines Reviewed," *Jour. Geol.*, Vol. XLVI (1938), pp. 985–95; Shepard, "Classification of Marine Shorelines: A Reply," *Jour. Geol.*, Vol. XLVI (1938), pp. 996–1006.

[2] "Revised Classification . . . . ", *op. cit.*, p. 617.

[3] *Op. cit.*, p. 991.

River and two of its tributaries, Fox Creek and Treadwell's Island Creek. The spit west of Steep Hill is particularly well shown. To the north of the Ipswich River are Little Neck and the southern end of Plum Island.

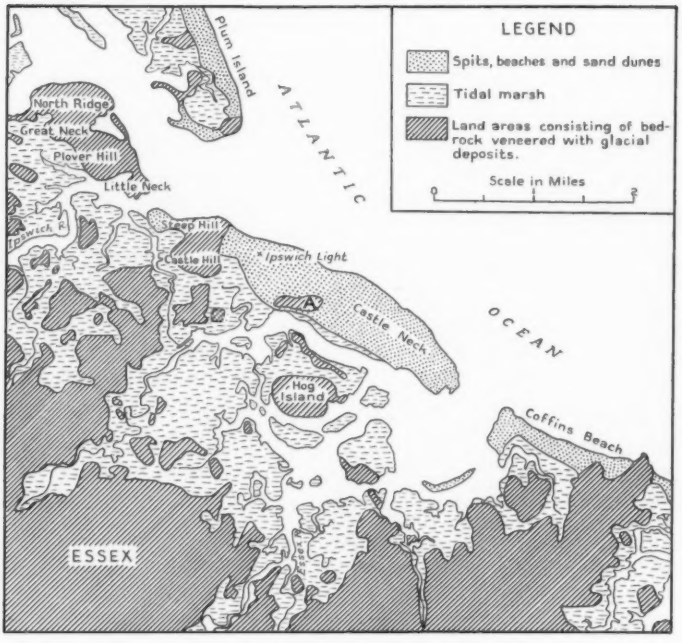

FIG. 1.—A map redrawn from part of the Salem and Gloucester topographic sheets, showing Castle Neck and the spit west of Steep Hill.

Shepard apparently believes that Castle Neck has grown eastward by the deposition of wind-blown sand, which could only be derived from an area west of the Neck. In support of this he mentions a new lighthouse which has been built beyond the old one because of dune obstruction. However, according to Captain George E. Eaton, of the United States Lighthouse Service, there has been only one lighthouse on Castle Neck. This was built in 1838 and remodeled in 1932. The movable range-finder that was situated for a short time northeast of the lighthouse, and which is shown on the chart reproduced in Shepard's figure, was removed in 1932. But

even if Shepard were correct and a lighthouse had been removed because of dune obstruction, this in itself would be no proof that Castle Neck had grown eastward because of migrating dunes. With the contention that Castle Neck is a wind-deposition coast in Shepard's sense the writers are not in agreement for the following reasons.

FIG. 2.—An aerial photograph of the spit west of Steep Hill. Photograph (1938) by Bradford Washburn, Institute of Geographical Exploration, Harvard University.

First, extending westward from the wave-cut cliff on the north side of Steep Hill there is an unnamed spit about $\frac{1}{2}$ mile long, which is covered with sand dunes, as shown in Figure 2. From a casual inspection of Figure 2, this area might be classified as a wind-deposition shoreline. However, field work shows that these dunes merely veneer a series of beach ridges which, as is indicated by the size of the boulders and pebbles in them, grew westward from Steep Hill. The dunes were formed of sand blown from these beaches. As Castle Neck is exactly the same in appearance as the spit under discussion, it does not seem illogical to conclude that Castle Neck was also formed as a series of beach ridges which were later blanketed with sand.

Second, the highest part of Castle Neck lies between 60 and 80 feet above sea-level. Viewed from a distance this area looks as if it were composed of many high dunes. However, field work shows that it is actually a small rounded hill of till veneered with dunes. This area is labeled *A* in Figure 1. It should be stated that, since its original deposition, the till has been modified by frost and wind action. It is in places covered by a typical warp containing wind-cut stones. This warp is similar to that found elsewhere in southeastern Massachusetts[4] and is sufficiently limonitized and weathered so that its features, particularly the wind-cut stones, need not be confused with any formed by present-day processes. The hill has a steep, straight cliff on the north side which is obviously an old sea cliff. The alinement of this cliff shows that the first beach which joined this hill to Steep Hill must have had a position much to the south of the present shoreline. No old beaches were found between this hill and Steep Hill, since the cover of dune sand is thick. There are, however, scattered beach pebbles in many places which indicate that the northwestern part of Castle Neck consists of a till hill which is tied to Steep Hill by a tombolo. The southeastern part of the Neck consists of a succession of dune ridges roughly paralleling the shore, irregular hillock dunes, and blowouts. No beach ridges have been found, but scattered beach pebbles and the alinement of the dunes indicate that first a spit grew southeastward from the hill under the prevailing shore currents and that thereafter it was prograded northward until the beach stretched from Steep Hill southeastward leaving the till hill inland. Further proof that the greater part of the material in the Neck was undoubtedly deposited by waves and shore currents lies in the character of the present beach which is pebbly during the winter months. Such a beach could not have resulted from the shaping of advancing dunes by waves and currents as suggested by Shepard.

Third, as Lucke[5] suggested in his review, there are theoretical conceptions which are strongly against classifying Castle Neck as a

[4] Kirk Bryan, "Geologic Features in New England Ground Water Supply," *Jour. New England Water Works Assoc.*, Vol. L (1936), pp. 222–28; see also Kirk Bryan, "The New England Ground Water Supply," *Harvard Alumni Bull.*, Vol. XXXIX (1937), pp. 676–81.

[5] *Op. cit.*, p. 991.

wind-deposition shoreline. A possible source for the sand, if we accept Shepard's theory for its origin, would be Steep Hill, Castle Hill, and contiguous areas of glacial material northwest of the Neck. This is an inadequate source as proved by the following facts: (1) it is obvious that these areas were always too small to supply all the sand necessary; (2) as these areas are composed largely of till, a thick boulder pavement should now cover them if any considerable quantity of sand had been carried from them by the wind: no such boulder pavement exists; (3) since the glacial period and during the time of beach formation, these hills were covered with vegetation, making them unavailable as a source of sand. If these glacial deposits are an inadequate source, the only other possible source would be the beach in front of Steep Hill. As the sand in this beach, throughout its existence, has been derived in the main from the eroding cliff cut on the north side of Steep Hill, it is obvious that from lack of quantity this, too, is an inadequate source.

Fourth, the wind rose for Boston indicates that the prevailing winds of the region are northwesterly. The dune forms also indicate that the effective wind is northwesterly. If Castle Neck had been formed because of wind deposition by the prevailing wind, one would expect it to be oriented not in the actual trend nearly W.NW. and E.SE. but rather northwest and southeast.

It is evident that the dunes have migrated southwestward so as to widen the Neck at the expense of the marsh. Moreover, the writers would agree that some sand is being blown off the dunes and beach and carried southeastward, thus extending the Neck. However, they feel that this process has been of inconsiderable importance.

On the grounds set forth above, the writers regard Castle Neck as a spit built by waves and currents and not as a wind-deposition coast in Shepard's sense. They do not deny the possibility that wind-deposition shorelines may exist. They believe, however, that as a class they must be rare, necessitating as they do an area back of the coastline capable of supplying large quantities of sand, a wind system which will carry this sand to the coast, weak waves and shore currents, and in all probability a special configuration of the coast.

DISCUSSION: EXAMPLE OF WIND-DEPOSITION SHORELINE

FRANCIS P. SHEPARD
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The foregoing criticism of one of the examples in the writer's classes of shorelines appears to be convincing. Evidently, the information given to the writer twenty years ago to the effect that the

FIG. 1.—An example of a "wind-deposition" shoreline. Photograph taken at the northwest end of the outer coast of San Clemente Island. Note the wave nip at the base of the sand dune. The beach in the foreground is exposed by the low tide.

lighthouse on Castle Neck had to be moved because of dunes was erroneous. The writer would further agree with Bryan and Nichols that this item in the classification is not a very important one, although it would appear to be necessary for completeness. In seeking examples of all of the twenty-four classes of shoreline, considerable difficulty was experienced so that it is not unlikely that others of the

illustrations may prove erroneous. Also certain of the classes may be shown to be unwarranted.

Since Castle Neck appears to be principally a shoreline of marine deposition and only to a minor degree prograded by the wind, a better example of a "wind-deposition" shoreline is required. Fortunately, a clear-cut example was discovered last year during a visit to San Clemente Island off the California coast. On this island the strong northwest winds have swept sand across from the beaches on the northwest side, and the resulting dunes are prograding the coast in small bites on the adjacent portions of the long southwest-facing coast. The accompanying illustration (Fig. 1) shows the best example of a prograding dune. The wave nip at the base shows how the waves reach the dune fore slope at high tide. No claim is made that the shoreline has been greatly extended.

Possibly better examples of "wind-deposition" shorelines may be found along the west coast of Africa, where quantities of wind-blown material are being carried out to sea from the Sahara Desert. Another locality which would appear to be favorable is the west coast of Australia. In these localities the prograding may at least locally exceed the wearing-back by the waves.

ARTHUR PHILEMON COLEMAN

1852–1939

A great geologist has finished his work. Arthur Philemon Coleman stands out prominently as one of the builders of our science, various parts of which were fashioned in an important way by his workmanship. Brought up near the southern border of the Canadian shield, his early work was chiefly with the pre-Cambrian rocks, structures, and ore deposits. His efforts were directed along many different lines, and the accomplished results were of prime importance. His notable investigations of the now-famous Sudbury nickel area, in particular, have called forth high tribute from geologists and mine operators alike. Leading to another branch of the science was his discovery of glacial tillite at the base of the Cobalt series, which attracted the attention of geologists the world over and started him on a long succession of critical studies of the ancient tillites on all the continents of the globe, except Antarctica. These and other glacial researches made Professor Coleman pre-eminent in this phase of geology.

He was one of the world's most widely traveled geologists, with no waning of enthusiasm with advancing years. Even after passing the fourscore mark, he utilized our northern winters for field work in the high Andes of Colombia and the mountains of Middle America to round out his studies of Pleistocene glaciation within the tropics; the summers found him back at work on the glacial features of southern Ontario. At eighty-six he was planning a journey to British Guiana, but unfortunately that was not to be.

Professor Coleman will be remembered not only as a scientist of the highest rank but also as a master of style in writing and lecturing, a painter of beautiful landscapes which have given pleasure to many friends, a pioneer mountain climber, honored by the Alpine Club of his own country and those of several other countries, and, above all, as one of the most lovable of men. His memory will long be cherished.

ROLLIN T. CHAMBERLIN

REVIEWS

Manual of Sedimentary Petrography. By W. C. KRUMBEIN and F. J. PETTIJOHN. New York and London: D. Appleton–Century Co., 1939. Pp. xiv+549; figs. 265. $6.50.

In this splendid compilation of sedimentary petrographic techniques the authors have collected and integrated a vast amount of data, much of which was gathered from the literature devoted to soil science, ceramics, hydrology, physics, and statistical methods. It is the first adequate handbook of sedimentary petrography published in the United States. The volume covers methods for every step of an analytical study of a sediment, from the field sampling to the final statistical or graphic representation of the analysis.

The book is divided into two parts of approximately equal length, each author assuming the responsibility for his portion. It is fortunate for the reader that each author has an excellent background of experience in his particular field. This gives assurance of technical competency in the handling of the subject matter of each portion of the book.

Part I, including chapters i–x, deals with sampling, preparation for analysis, mechanical analysis, and statistical analysis. A chapter is devoted to each of the following subjects: definitions and properties of grains (8 pages), collection of samples (32 pages), preparation of samples for analysis (33 pages), concept of grade size (15 pages), principles of mechanical analysis (44 pages), methods of mechanical analysis (47 pages), graphic presentation of analytical data (30 pages), elements of statistical analysis (16 pages), application of statistical measures to sediments (40 pages), and orientation analysis of sedimentary particles (9 pages).

Part II, including chapters xi–xxi, deals with shape analysis, mineralogical analysis, chemical analysis and mass properties of sediments. A chapter is devoted to each of the following subjects: shape and roundness of fragments (26 pages), surface textures of fragments (6 pages), preparation for mineral analysis (10 pages), separation methods (38 pages), mounting for microscopic study (9 pages), optical methods of identification of minerals (46 pages), description of minerals of sedimentary rocks (53 pages), chemical methods of study (8 pages), mass properties of sedi-

ments (24 pages), and laboratory equipment and organization of work (11 pages).

Most of the foregoing units are presented with a reasonable balance between theoretical aspects and practical application. Exceptions, perhaps, are the chapters on statistical methods and their application and the discussions pertaining to the distinction between roundness and sphericity of mineral grains. These units seem to be overemphasized, for only a very small minority of students of sedimentation are interested in the mathematical theories involved.

Since the volume was intended to serve as a manual for laboratory work, the reviewer feels that there is a lack of summaries or flow sheets which give the essential points of procedure for the more routine methods employed in laboratory work. Diagrams such as Figures 18, 68, 264, and others, are of great aid to students. Likewise, the descriptions of minerals and rocks in chapter xvii deserve special commendation. The concise descriptions and tabulations of diagnostic properties of detrital mineral grains, together with the numerous well-executed sketches, facilitate the identification of unknown mineral fragments. The chapter on graphic presentation of analytical data gives an equally concise and well-illustrated treatment.

Many first editions contain errors, and this volume is no exception. However, the number noted by the reviewer is not great, and the errors are so obvious that the reader is not misled seriously. The description of Figure 102 in the second paragraph on page 217 and the omission of letters on Figure 170, page 342, are examples. Similarly, several mathematical errors in simple division were observed.

The book is so packed with analytical data that it may serve as a handbook not only for sedimentationists but also for industrial and research engineers who have to deal with fragmental materials. Its style is compact and its discussions are closely interwoven with graphs and diagrams. Numerous footnote references direct the reader to the literature where the various methods were first described.

The authors are to be congratulated on the preparation of a book which is virtually a one-volume reference library on sedimentary techniques and which will assist in the attainment of a higher degree of accuracy in problems of sedimentary petrographic analysis. It is indispensable to research in sedimentation.

GEORGE A. THIEL

Santorin: Der Werdegang eines Inselvulkans und sein Ausbruch, 1925–1928, Vols. I–III. By HANS RECK, assisted by FRITZ BEHREND, FRIEDRICH DOBE, GEORG GEORGALAS, DEMETRIUS HONDOS, NIKALAUS LIATSIKAS, M. N. VAN PADANG, WERNER QUENSTEDT, and JULIUS SCHUSTER. Berlin: Dietrich Reimer, 1936. Vol. I, pp. 187+xvi; figs. 1–34; colored plates, 2; Vol. II, pp. 353+xv; figs. 35–109; colored plates 2; Vol. III, pp. xiii; pls. 101.

These three beautifully prepared volumes give an excellent and detailed account of recent volcanic activity in the great caldera of Santorin. The Cyclades and Sporades Archipelago between Greece and Asia Minor may be regarded as two concentric arcs. The inner or northern arc is made up chiefly of crystalline rocks; the outer arc, embracing Aegina, Milos, Santorin, Nisyros, and Kos, exhibits chiefly young volcanics, marked by fracture zones which, according to the authors, pass radially from the center of the arcs through the several islands mentioned and appear to have served as passageways for the lavas.

The Santorin group itself is roughly circular and consists today of two major islands which outline the famous caldera. The eastern and larger, Thera, or Santorin Island, is crescentic, 15 by 9 kilometers in size, and made up of tuffs and lavas, with small remains of schists and limestone between. The northwesterly island, Therasia, is smaller, measuring 6 by 3 kilometers. To the southwest the caldera leads into the Aegean through a large opening. Previous to 1925 a group of three islets, designated from southwest to northeast, Palaea Kameni, Nea Kameni, and Mikra Kameni, occupied the approximate center of the sea-filled caldera, which averages 8 kilometers in diameter.

The caldera itself presents a problem, reminiscent to Americans of that of Crater Lake (the extensive literature of which was apparently unknown to the authors). Santorin basin was regarded as an explosion crater of gigantic size by von Seebach, von Fritsch, and Washington, and tentatively accepted as such by Fouqué, though with reasoned misgivings. After a study, partly of the pumice cover of the peripheral islands, partly of the radial fracture zones which he clearly recognized, Professor Reck assigns the caldera to collapse.

The general geologic setting of the Cyclades, the origin of typical calderas, and the description of the history of the Santorin caldera occupy the greater part of Volume I. This section, if slightly prolix, unquestionably presents an interesting argument, carefully developed.

The second volume deals primarily with the recent eruptions on the Kameni Islands, the first at Santorin since that of 1866 to which the

Kameni owe their presence. Originating in the deep but narrow channel between Nea Kameni and Mikra Kameni and presaged by a great submarine explosion and shock on July 28, 1925, came repeated steam eruptions, with concomitant reddening of the water, and by August 12 the cone of Daphni had risen above the sea. Immediately Professors Georgalas and Liatsikas of the Greek Geological Survey hurried to the scene, being later joined by the German and Dutch members of the group and the Greek physicist Professor Hondos. Eventually the two great lava flows from Daphni filled the channel between the bordering islands, uniting Mikra and Nea Kameni into one. The lavas and gas eruptions came from a main vent the position of which shifted progressively southwestward, apparently along a definite fissure, as determined by observations of the steam cloud made from Thera. Two main periods of eruption occurred, ending in vigorous explosions and the formation of a lesser cone ("Nauplius"), which largely covered the earlier features with ash.

Among the gases, SO_2 was prominent in earlier stages, H_2S in later ones, while HCl was detected throughout. The cauliflower and "incandescent" clouds, like those so destructive on Martinique, as well as the electric discharges and the "trombes" (volcanic tornadoes), were subjects of special study. So also among the flow phenomena were the relations between lateral movements of lava and accessory cones and the more or less horizontal gas ejections, as well as the sequence and relationships of the great "primary" flows and the "secondary" ones, representing outbreaks of lava through the crusts of the older flows. Of special interest too is the large amount of iron carbonate which was discharged into the sea by hot springs near the eruption; when this rose to the surface of the water, it became oxidized and sank again to form a thick cover of iron hydroxide, reminiscent of Van Hise's earlier explanations of those Lake Superior iron ores associated with greenstones. In one locality the iron content of these waters was observed to give way to sulphur as H_2S began to be emitted.

Space does not allow a more detailed discussion of this excellent report. The illustrations in the third volume are masterpieces of photography, emphasizing especially the use of the camera in recording progressive changes. Four color plates greatly increase the attractiveness of the publication. Of greatest value is the accurate description of the advance of the lava and the changes in form of the volcano corresponding to the various stages of the eruption.

CHARLES H. BEHRE, JR.

Permian Trilobites from Timor and Sicily with a Revision of Their Nomenclature and Classification. By R. F. C. R. GHEYSELINCK. Amsterdam: Scheltema & Holkema's Boekhandel, 1937. Pp. 108; pls. 4; figs. 31.

Gheyselinck presents a conservative view of generic classification in the latest of recent important papers on late Paleozoic trilobites. His conclusions contrast with those leading to the closer generic distinctions of Toumansky, Weber, and Weller, and his interpretations of generic affinities differ sharply from those of Toumansky and of Weller. The accompanying table summarizes the generic and subgeneric relationships presented:

Genus Brachymetopus McCoy. Cheiropyge Diener(?)
  Subgenus Brachymetopina Reed—not established
Genus Proetus Steininger
  Subgenus Permoproetus Toumansky
Genus Phillipsia Portlock
  Subgenus Phillipsia *sensu stricto*. Ameura Weller(?)
  Subgenus Neogriffithides Toumansky—considered transitional to Griffithides *sensu stricto*
  Subgenus Paraphillipsia Toumansky—considered doubtfully worthy of subgeneric standing
  Subgenus Neophillipsia nov. Sevillia Weller pars.
Genus Griffithides Portlock
  Subgenus Griffithides *sensu stricto*
    Group *a* Exochops Weller
    Group *b* Neoproetus Tesch
    Group *c* Kaskia Weller, Paladin Weller
  Subgenus Pseudophillipsia Gemmellaro. Anisopyge Girty, Ditomopyge Newell, Cyphinium Weber, Sevillia Weller *sensu stricto*

Four genera are considered valid, five are reduced to subgenera, and thirteen are placed in synonymy or are considered doubtful.

The material studied consists of collections from the Permian of Timor and of Sicily. Nine new species and varieties from Timor and one new species from Sicily are described. The new subgenus and species *Neophillipsia decurtata* is based upon a specimen from the Pennsylvanian of Kansas. A new name, *Griffithides welleri*, is given for *Griffithides transilis* Weber varieties *alpha* and *beta*.

The author writes in English of passable clarity for the most part, although misspellings and use of inappropriate words obscure the meaning of some sentences. The numerous text figures are valuable; the plates illustrating new species are only fair. Gheyselinck's statement (p. 49) that the genus *Anisopyge* Girty was founded upon *A. antiqua* as well as upon

Phillipsia perannulata Shumard is in error, since Girty clearly questioned the generic assignment of *A. antiqua*.

The invariable statement of textbooks that Permian trilobites are few in number and belong to a handful of species is distinctly out of date. The reviewer has found in the literature names and descriptions of sixty-eight species and varieties. This number rivals, if it does not exceed, that known from the Upper Carboniferous. In fact, the work of the last thirty years has added so much to our knowledge of Permian faunas that the total number of invertebrate species known from Permian rocks is much greater than the total number known from Upper Carboniferous strata.

C. C. BRANSON

Contributions to the Knowledge of the Chemical Composition of the Earth's Crust in the East Indian Archipelago, Part I: *The Spectrographic Determination of the Elements according to Arc Methods in the Range 3600–5000 A;* Part II: *On the Occurrence of Rarer Elements in the Netherlands East Indies*. By W. VAN TONGEREN. Amsterdam: D. B. Centen's Uitgevers-Maatschappij, N.V., 1938. Pp. x+181; figs. 14. Paper, bound together, Dutch fl. 4.50; Part I alone, Dutch fl. 3.00.

Part I is essentially a laboratory manual for spectrographic mineral analysis. It contains (1) a description of the apparatus and equipment used, (2) a brief discussion of spectrographic technique, and (3) a detailed study of the spectrographic behavior of the elements. The apparatus and methods are conventional in this type of analysis. A glass prism spectrograph is used. Excitation is by the d-c arc, with carbon electrodes, using the Glimmschicht method, with the sample in the negative electrode. A 10-mg. sample is used for the analysis. The sample, with an equal weight of sodium carbonate, is placed in the crater of the lower electrode and arced until all the material is vaporized. During the arcing, fractional exposures, each of 30 seconds' duration, are made.

The detailed study of the spectrographic behavior of the elements is carefully done. The elements are individually considered. For each there is given a discussion of the spectrographic sensitivity, the advisability of this method, a tabulation of sensitive lines, and a list showing the order of disappearance of lines with decreasing concentration. Following this section there is a list of analysis lines, arranged by wave-lengths, showing the relative intensities of lines, the intensities at various concentrations in a silica base sample, and the interfering substances. Finally, the author gives complete directions for preparation of a standard analysis powder with a silica base and forty-two elements. The analysis lines given by the

standard mixture are listed in order of wave-length. The described analysis powder is similar in principle to the R.U. powder marketed by Hilger, but it promises to be more useful to the mineral analyst because of the known concentrations and the use of a silica base.

Part I is a valuable contribution for the spectrographic analyst. The individual discussion of each element will enable the analyst with no previous experience in the arcing of mineral samples to select best conditions and to evaluate the sensitivity to be expected from each element. It is unfortunate that the author's equipment did not include a quartz spectrograph, since many of the elements can best be determined in the wave-length region below that transmitted by glass. The user of a quartz instrument will need to extend the tables to cover the range below the glass instrument.

Part II contains the results of the author's analyses on some three hundred samples. Complete descriptions are given of the source and preparation of the samples. Analytical results are tabulated according to the geographical origin of the sample. Semiquantitative estimates, based on the averages of several determinations, are given for each constituent identified. In a summarizing discussion the the author briefly considers the geochemical importance of each element and discusses the occurrence of the element in his samples.

Although the analytical results are limited by the wave-length region and by the geographical source of the samples, the completeness of the author's analysis will render his findings of value. The accumulation of data such as these is a laborious task, and the author is to be praised for his painstaking and careful work.

W. C. PIERCE

Miocene Stratigraphy of California. By ROBERT M. KLEINPELL. Tulsa: American Association of Petroleum Geologists, 1938. Pp. 450; pls. 22; pocket charts 5. $5.00 ($4.50 to A.A.P.G. members and associates).

This volume deals with the Miocene stratigraphy of the Reliz Canyon section of Salinas Valley, Monterey County, California. In this locality about 6,000 feet of Monterey shale separate the Vaqueros and Santa Margarita sandstones. Foraminifera collected from this shale furnish a basis for the stratigraphical studies.

The foraminifera represented in Reliz Canyon have been grouped, according to Cushman's classification, into ten families and thirty-five genera. They are distributed through the lower half of the Monterey shale where they form the basis of sixteen minor biostratigraphic units.

Two such units were also recognized from shale partings in the upper Vaqueros Formation. Ecologically, the foraminifera represent temperate-zone assemblages, comprising six distinct faunal facies. With very few exceptions, the faunules suggest deposition at varying depths between the upper neritic zone and the edge of the continental shelf, i.e., between 25–500 fathoms.

The Monterey shale is considered to be of Miocene age in the sense used in Pacific Coast stratigraphy. Correlation of the Reliz Canyon section with the typical European column cannot yet be made accurately since the stratigraphic ranges of most of the foraminiferal species are not fully known.

A large section of the volume is devoted to a systematic catalogue and description of foraminiferal species which have been recorded from the California Miocene in the present and previous publications. In addition to many well-executed plates, a pocket contains several stratigraphic, ecologic, and correlation charts.

RAYMOND E. JANSSEN

The Gold Missus: A Woman Prospector in Sierra Leone. By KATHERINE FOWLER-LUNN. New York: W. W. Norton & Co., Inc., 1938. Pp. 303.

An interesting narrative of the author's exploration in search of mineral deposits in equatorial Africa, recording the discovery of iron ores, molybdenum ores, and gold, both lode and placer, with a popular description of their geologic setting and methods of prospecting.

N. L. B.

"Outline of the Geology and Mineral Resources of Goochland County, Virginia," by CARL B. BROWN. *Virginia Geological Survey Bulletin 48* ("County Series," No. 1). University, Va., 1937. Pp. 68; pls. 10; tables 5.

This publication, written especially for teachers and high-school students, discusses the geography, geology, mineral resources, and geologic history of Goochland County, Virginia. The county lies along the north bank of the James River in the Piedmont province, a short distance west of Richmond. In addition to photographs and tables, the booklet includes a glossary of geologic terms and a geologic map showing the locations of all mines and quarries within the county.

RAYMOND E. JANSSEN

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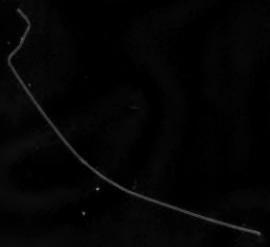
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